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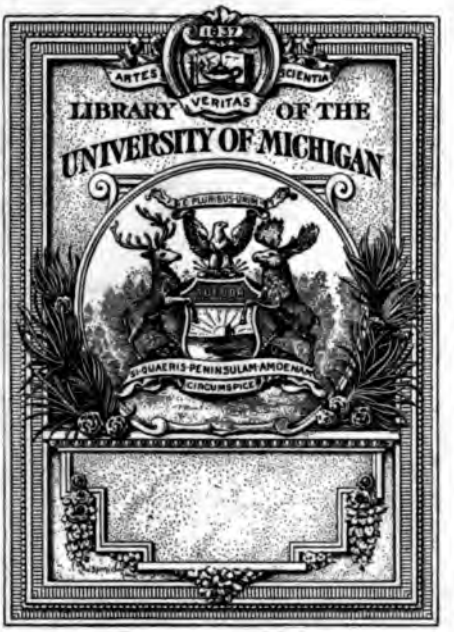
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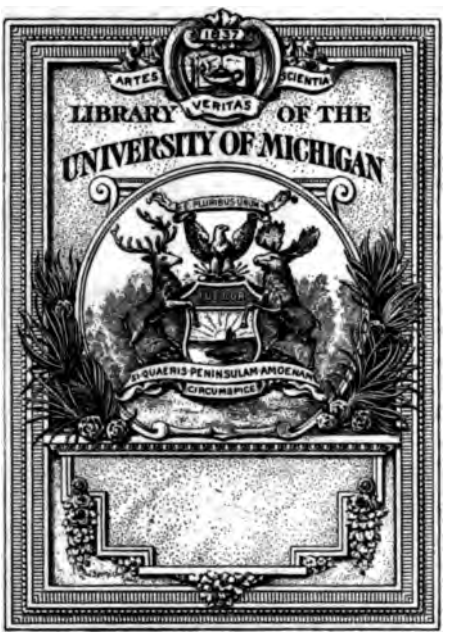
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METAMORPHIC GEOLOGY

A TEXT-BOOK

BY
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AND
W. J. MEAD
UNIVERSITY OF WISCONSIN

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INTRODUCTION

Rock metamorphism is here defined to cover all mineralogic, chemical, and physical changes in rocks subsequent to their primary crystallization from magma. Metamorphism has often been restricted to the changes involved in the development of slates, schists, gneisses, and associated crystalline rocks, as illustrated by the common textbook classification of rocks as *igneous*, *sedimentary*, and *metamorphic*. We shall follow Van Hise by including under metamorphism not only development of schistose and crystalline rocks, but also all changes involved in rock weathering and cementation.

Much of the literature on metamorphism consists of more or less unrelated detailed descriptions of rock alterations, but there are, in addition, a number of well known and valuable general treatises on metamorphism or some of its phases, which are indispensible to the student of the subject. Three of these, perhaps best known and most accessible to American students, are Merrill's discussion of rock weathering,[1] Van Hise's monograph on metamorphism,[2] covering the entire field from physical, chemical, and mineralogical standpoints, Clarke's bulletin on geochemistry,[3] covering the field principally from the geochemical standpoint. These discussions and others of a similar nature have been prepared primarily for the professional investigator. Experience in teaching the subject, both in class room and field, has seemed to show the need of a general text, eliminating many details, presenting the current knowledge in some perspective, and particularly indicating methods of quantitative approach to the subject which have in recent years been found useful in inductive field and laboratory study. This text is an attempt to meet

[1] Merrill, G. P., Rocks, rock weathering and soils: New York, 1906.

[2] Van Hise, C. R., A treatise on metamorphism: Mon. 47, U. S. Geol. Survey, 1904.

[3] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911.

this need. It is made purposely brief, as a guide to the study of principles as inferred from field and laboratory study. It is not a handbook of metamorphism in which one may expect to find an adequate description of metamorphic details. Certain new features are presented which have arisen in our own investigations. We hope that it may be found useful to others interested in this phase of geology, and that geologists not before familiar with metamorphism in some of its larger aspects may find help in it. Almost any phase of geology, when systematically developed, may afford a convenient means of approach to a wide range of geologic problems. Our treatment of metamorphism in the following text is influenced to some extent by this consideration, for we find in the subject of rock metamorphism a convenient basis of approach to a considerable variety of geologic problems.

Throughout the following discussion an effort is made to express metamorphic changes quantitatively with a view of bringing them into perspective. In order to do this without using too many figures, groups of actually measured changes have been summarized or averaged and the discussion is based on the averages. No finality is claimed for these averages,—in studying individual cases of metamorphism wide departures from the average are to be looked for,—but the average cases are convenient standards to start with in defining the changes common to certain groups of metamorphic rocks.

ACKNOWLEDGMENT

We have drawn freely on the available literature on metamorphism. In most cases credit is given in the text but no attempt has been made to make the bibliography complete, for the book is not written for the investigator following up the subject in detail. Clarke's bulletin on geochemistry is a most satisfactory reference book as an accurate summary of facts of geochemistry, with comparatively full reference to the literature. Our primary obligation, however, is to the work of C. R. Van Hise, who was really the first to cover the field of metamorphism broadly and systematically. The senior author had the good fortune to work with Doctor Van Hise during the preparation of his monograph on metamorphism, and since that time, in association with the

junior author, has been following the subject from a quantitative standpoint with a view of testing and checking the conclusions of that treatise. We are indebted also to members of the Geological Department of the University of Wisconsin for aid and criticism.

CLASSIFICATION OF METAMORPHISM

Metamorphic processes may be divided broadly into two great groups:[1] *katamorphic*, or destructive alterations, as illustrated by weathering, and *anamorphic*, or constructive alterations, as illustrated by the development of schistose and crystalline rocks from unconsolidated sediments by rock flowage or by igneous metamorphism. The term "katamorphism" is used to cover all alterations of a disintegrating or decomposing nature, whether accomplished by weathering or by thermal solutions, at the surface or below. The term "anamorphism" is used to cover all integrating processes, whether by dynamic metamorphism or rock flowage, by cementation, by metasomatic replacement, or under the influence of igneous rocks,—wherever these changes may occur, near the surface or at great depth. Where convenient, the different types and conditions of katamorphism and anamorphism will be discriminated in describing the rock alterations.

For reasons which will appear later, we do not follow Van Hise in his emphasis on *zones* of katamorphism and anamorphism as controlling factors in classification, but rather devote our attention to an inductive study of the results and processes of metamorphism in terms of rocks, minerals, and chemical composition, regardless of the locus of the changes. Each rock is followed through its cycle of destructive and constructive changes. In short, the *metamorphic cycle*, described hereafter, is used as a basis for our approach to the subject.

THE METAMORPHIC CYCLE

The keynote of rock alterations is adaptation to environment. A molten rock or magma enters the outer shell of the earth and comes within our range of observation. No sooner does it crystallize than changes begin—with great rapidity nearest the surface,

[1] Van Hise, C. R., A treatise on metamorphism: Mon. 47, U. S. Geol. Survey, 1904, pp. 159 et seq.

with less rapidity below. These changes are both chemical and physical. Water, carbon dioxide, oxygen, and other substances of the hydrosphere and atmosphere attack the rocks. The ferrous iron of the igneous rocks combines with oxygen and water and a large part of the iron remains as residual limonite. Alkalies and alkaline earths form soluble salts and are leached out in well-determined order. Free quartz is less readily changed or dissolved. The portions of the bases which the waters have failed to abstract tend to remain combined with alumina and silica, taking on water to form new hydrous silicates, although some of them may remain in place as carbonates or other salts. The excess of alumina and silica left after the leaching of the bases from the silicates becomes hydrated and forms clay. There are ultimately left, then, iron oxide, quartz, clay, and a variety of hydrous aluminum silicates, characterized by a lower ratio of silica to the bases than in the original silicate minerals. Oxide zones of ore deposits are special cases of these residual products. These substances on the erosion surface are mechanically distributed and ultimately become segregated as mud and sand, or even iron ores. The substances which are taken out in solution may remain in solution in the sea or may be redeposited as limestone, dolomite, chert, iron carbonate, or iron silicate—in other words, as chemical sediments. Part of these substances also may be deposited as cements. The ultimate results of the destruction of the original igneous rocks, then, are the sediments, cements, salts of ocean, river, lake, and ground waters, and the un-transported residual products of rock decay.

The part of this mantle of products of rock weathering remaining on the continents has been called the *regolith*.[1] The subdivisions of the regolith, according to Merrill, are as follows:

[1] Merrill, G. P., Rocks, rock weathering, and soils: 1906, p. 287.

The regolith.	Sedentary.	Residual deposits.	Residuary gravels, sands, and clays, wacke, laterite, terra rossa, etc.
		Cumulose deposits.	Peat, muck, and swamp soils, in part.
	Transported.	Colluvial deposits.	Talus and cliff debris, material of avalanches.
		Alluvial deposits (including aqueo-glacial).	Modern alluvium, marsh and swamp (paludal) deposits, the Champlain clays, loess, and adobe, in part.
		Æolian deposits.	Wind-blown material, sand dunes, adobe and loess, in part.
		Glacial deposits.	Morainal material, drumlins, eskers, osars, etc.

The materials of the regolith ultimately may be transported and deposited as sediments in the ocean. The *regolith*, referring as it does to place, contains only a part of the end-products of weathering.

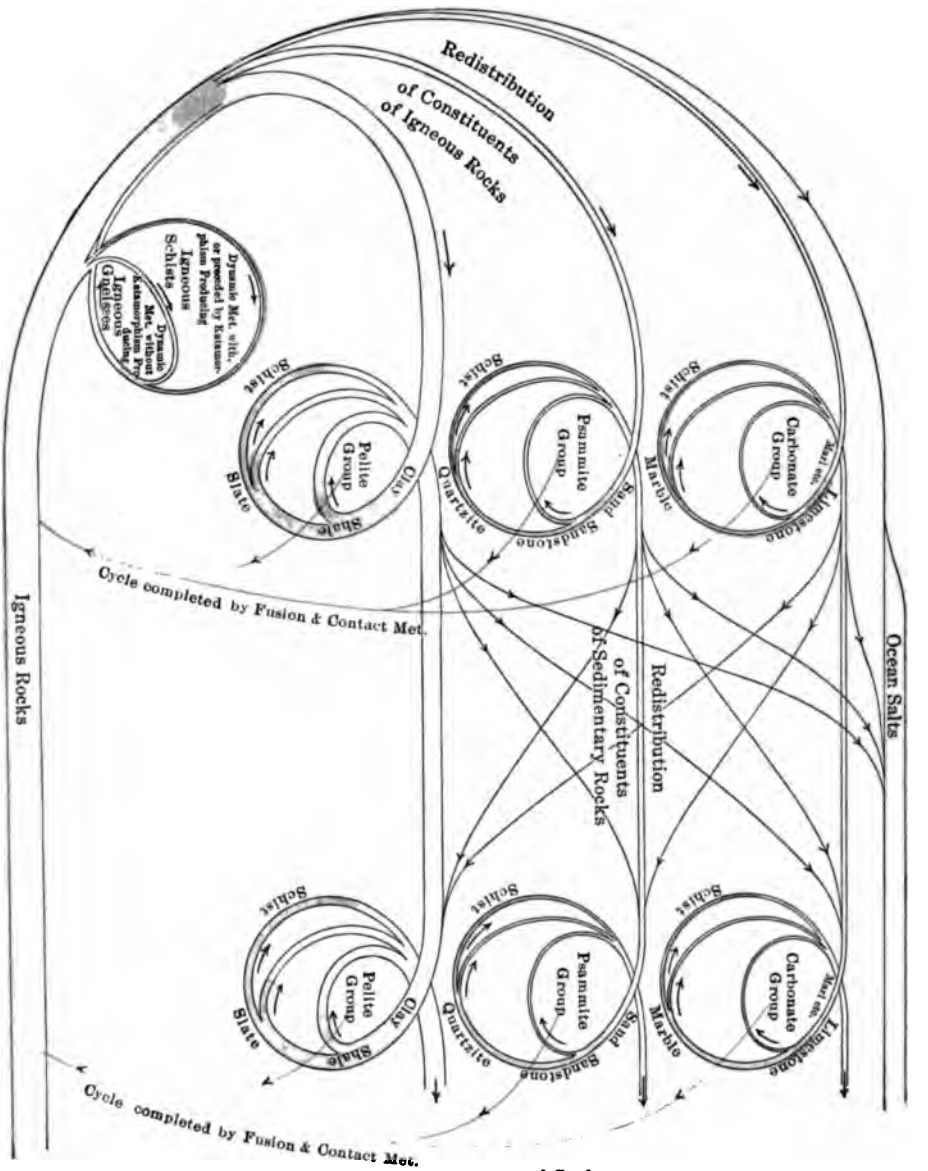
The extreme upper portion of the regolith, that which affords food and foothold for plant life, is commonly designated as soil. This brings in another factor, the capacity of the regolith for plant growth, and this requires classification of a different sort, with which we are not primarily concerned.

The katamorphic changes above outlined involve the addition of substances of light molecular weight, such as carbon dioxide, water, and oxygen. In terms of elements oxygen is of dominant importance. The resulting minerals are on the whole of simpler molecular composition and of lower specific gravity. The actual addition of new substances has also increased volume. The unconsolidated products have a large pore space which still further increases the volume. Energy is released in an enormous amount. Hot waters associated with igneous rocks locally expedite these changes.

This change was called katamorphism by Van Hise. By katamorphism the rock tends to become adapted to a surface environment. The resulting product may be compared to a solution, both solid and liquid, of the original rock substances with the atmosphere and hydrosphere.

But there is ever present a reverse or anamorphic tendency which has greater effectiveness in producing results in proportion as the substances on which it works have suffered katamorphism. Its effects on the sediments are conspicuous. On igneous or crystalline rocks they are much less conspicuous. No sooner have sediments been deposited than they tend to become consolidated by settling and by infiltration of cements. This process of induration is aided by the pressure of the accumulating sediments, and as the pressure due to burial or earth movements increases, other changes take place. Where the rock yields by movement, a schistose structure may be developed. The substances which had been added in katamorphism are forced out: first the water, then carbon dioxide, ultimately oxygen. The bases are recombined with the alumina and silica to reproduce some of the silicates of the original rock; not all of the silicates, however, for certain of the bases have been carried to the sea where they remain in solution, and thus the original ratios of bases to alumina and silica cannot be reproduced without the addition of bases from without. Feldspars, which constitute 50 per cent of average igneous rocks, are reproduced in average schists to a much less extent. The mineral molecules are on the whole more complex. Volume decreases by elimination of pore space, by elimination of substances, by new molecular groupings of higher density. Energy is absorbed by the new molecular combinations. In proportion as the rocks are buried beneath the surface or suffer mechanical deformation or contact metamorphism, these anamorphic changes are expedited. Anamorphism tends in general to make the rocks approach igneous rocks in texture, minerals, and chemical composition, but so far as the end-products are schists, gneisses and slates, they differ in certain essential respects from igneous rocks, mainly in the smaller proportions of feldspar, and augite, and larger proportions of mica, hornblende, and chlorite. The controlling factor is inferred to be the changes in the direction of producing hornblende, mica, and chlorite, so characteristic of the schistose and gneissic rocks. There is evidence locally along igneous contacts of actual fusion, in which case igneous rocks are reproduced and the cycle is complete. In such cases materials from the igneous rock have permeated the sediments and supplied the certain deficiencies in composition

PLATE I



Graphic summary of the metamorphic cycle, illustrating the changes in rocks in their progress through the various phases of metamorphism. Katamorphic alterations are represented in red and anamorphic alterations in green.

caused by katamorphism. Whatever the probability of fusion on a large scale, it is difficult to prove by inductive reasoning from field evidence.

The succession of katamorphic and anamorphic changes above outlined constitutes the metamorphic cycle. It is not necessary to assume that all rocks undergo this cycle. The cycle may be changed to a reverse phase at any point. The recurrence of cycles of the type above described are in a certain sense pulsational or oscillatory. Van Hise has used the term "zone of katamorphism" for the region near the surface where katamorphic processes are at a maximum and the "zone of anamorphism" for a deeper region where the reverse changes are at a maximum; thus emphasizing influence of depth on metamorphic changes. These terms are extremely useful in the general sense in which Van Hise used them. It is difficult, however, to use them in the inductive study of rock changes, for so many other factors enter that at any depth, or at the same depth at different times, the changes may be in opposite directions for different kinds of rocks. The conception of the metamorphic cycle is better adapted to an inductive study of metamorphic rocks, for it relates merely to the succession of changes in the rock without reference to depth.

In Plate I an attempt has been made to represent graphically the progress of the principal classes of rocks through the metamorphic cycle. Each of the rock classes may go through minor cycles, and both major and minor cycles may be short circuited at almost any point. A general conception of this kind may be only crudely expressed by graphic methods, but it may aid in an understanding of the broader relationships of the different phases of metamorphism discussed in following chapters.

PART I

KATAMORPHISM

CHAPTER I

KATAMORPHISM OF ACID IGNEOUS ROCKS. WEATHERING OF GRANITE AS A TYPE OF KATAMORPHISM

Acid igneous rocks of the granitic type are usually regarded as the most abundant of the igneous rocks. In Clarke's [1] average analysis of the igneous rocks, the granite types are given predominance over the more basic types in a ratio of 3 to 1. (See also pp. 64–67.) A study of the alteration of a granite at the earth's surface should therefore disclose the essential processes and results of katamorphism. We may consider the katamorphism of a granite in terms of (a) rocks, (b) mineral alterations, (c) changes in chemical composition, (d) volume changes, and (e) energy changes.

The terms *katamorphism* and *weathering* will be used more or less interchangeably. Weathering is most effective near or at the surface, and involves changes which are typically katamorphic in nature.

KATAMORPHISM AND REDISTRIBUTION OF A GRANITE IN TERMS OF ROCKS

When a granite is altered primarily by disintegration, a grit or arkose is the result. Where chemical decomposition is important, as in humid warm climate, there is developed a residual clay or soil containing more or less quartz sand. On transportation, sorting, and redeposition, this separates into sand and clay. The substances extracted from the granite in solution may be precipitated, with or without the aid of organic agencies, as carbonate, or chert formations, as cements in fragmental formations, as adsorbed salts in soils and clays, or they may remain as salts in the ocean. The proportions of these products resulting from a

[1] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, pp. 24–27.

given case of katamorphism may be approximately calculated. (See Chapter VII, Part I.) Clay is the dominant end-product.

KATAMORPHISM OF A GRANITE IN TERMS OF MINERALS

Qualitative Consideration of Mineral Alteration of a Granite

The average mineral composition of granites may be taken as 52 per cent feldspar, 14 per cent ferromagnesian minerals and mica, and 31 per cent quartz. In the progress of katamorphism the feldspars and the ferromagnesian constituents lose their bases and, by hydration, become kaolin. In the absence of iron stains, this causes a whitening or bleaching of the rock. The iron in the ferromagnesian minerals under surface conditions may be oxidized in place, becoming hematite or limonite, giving yellowish and reddish colors to the altered product. Under reducing conditions part of the iron may be taken into solution, to be later precipitated as iron carbonate or hydrous oxides, commonly limonite. The silica leached from the rock is deposited as chert or quartz. The small amounts of magnesia and lime taken from the granite are principally deposited as calcite and dolomite, either in limestone or dolomite formations, or mixed with the fragmental formations. Soda largely goes to salt in the sea, although some sodium salts may be precipitated under appropriate conditions of desication. Potash remains largely with the clay in more or less doubtful combination. (See p. 88.) The ultimate mineral products of the alteration of a granite, therefore, are clay or kaolin, quartz, iron oxide, calcite, dolomite, and salts in the sea. Hydrous aluminum silicates are the dominant product of weathering. Under certain conditions clay is altered to aluminum hydrates and silica. In proportion as decomposition is incomplete, the original minerals of the granite remain mixed with these end-products, and under certain conditions, as in arid climates, these original minerals may be the principal minerals in the products of katamorphism. The ultimate mineral products are of simple composition and on an average of lower specific gravity than the original minerals.

While the above are the characteristic end-products of the weathering or katamorphism of a granite, there are intermediate minerals developed during decomposition. Sericite may form, with

clay, in the destruction of the potash feldspars. Lime-bearing feldspars may yield epidote and calcite. Augite may be changed to hornblende or chlorite or bastite. A fibrous form of hornblende, secondary to augite, is called *uralite*. The hornblende may alter to chloritic products. The ferromagnesian minerals may change to talc and serpentine. Biotite alters to chlorite and magnetite. Most of these changes go on, however, with the separation or elimination of one or more of the ultimate end-products of katamorphism, like carbonates, kaolin, or quartz, and may be regarded as steps in the process of complete decomposition.

Of the original minerals the pyroxenes and amphiboles yield most readily to weathering, then follow the plagioclase feldspars, then the orthoclase and micas, with muscovite the most resistant. Quartz is less soluble than these minerals but ultimately it yields to solution. Of the accessory minerals, apatite and pyrite are most easily decomposed and minerals like magnetite, zircon, corundum, chromite, etc., tend to accumulate with little alteration in the sands.

Quantitative Consideration of Mineral Changes in Katamorphism of Georgia Granite

For illustration we describe the alteration of a Georgia granite. This may be regarded as a fairly typical case of advanced katamorphism. The fresh rock is from about 15 feet from the surface and the decomposed material from a depth of about 5 feet. It is assumed that before decomposition the rock sampled was essentially uniform in chemical and mineralogical composition. These fresh and altered materials are described by Watson [1] as follows:

"The fresh granite is dark-gray in color, and medium fine-grained, massive in texture. . . . Microscopically, the component minerals are quartz, orthoclase with microperthitic intergrowths of a second feldspar (albite), microcline, plagioclase near oligoclase, biotite, a little muscovite, apatite, zircon and magnetite. The biotite is much altered to chlorite and epidote. The epidote is very abundant, as large grains with well defined crystal boundaries. The comparatively large amount of this mineral is shown in the large percentage of lime in the analysis. Some secondary muscovite is present, derived from the feldspathic alteration. . . .

[1] Watson, Thomas L., Granites and gneisses of Georgia: Bull 9–A, Geol. Survey of Georgia, 1902, pp. 310–311.

"This (the weathered rock taken 5 feet below the surface) is a deep red clay, which, owing mainly to the presence of free quartz and some of the undecomposed silicates, has a distinct gritty feeling. Upon separating the silt and clay from the coarser particles, by washing, an abundance of the biotite, fresh quartz and kaolinized feldspars are left as a residuary sand. The biotite is in various stages of decay, varying from the thoroughly leached brass-color to black, partially altered folia. The largest grains of the partially decayed minerals, of the residual sand, measure 2.5–3 mm. in diameter; while the bulk of these fall within the range of one millimeter. They vary, from this size down to the smallest silt and clay particles –0.1–0.005 mm. Removal of the iron sesqui-oxide staining from the clayey product by renewed digestions in hot dilute hydrochloric acid, showed a large proportion of fresh grains of biotite, feldspar and quartz, with sharp, angular borders."

The mineralogical compositions of fresh and altered granite, calculated from the chemical analyses on page 8, are given in Table I. For explanation of method of calculation, see pages 308–309.

TABLE 1. MINERAL CHANGES IN KATAMORPHISM OF GEORGIA GRANITE ON ASSUMPTION THAT ALUMINA REMAINS CONSTANT

	Fresh granite. Grams per 100.	Altered granite. Grams per 100.	Altered granite. Grams per 100 grams of fresh rock. (Al_2O_3 assumed constant.)
Biotite	5.59		
Orthoclase molecule	16.60	14.30	9.65
Albite "	36.90	16.20	10.94
Anorthite "	16.20	3.73	2.52
Kaolin	1.39	50.30 [1]	33.95
Quartz	22.41	12.77	8.62
Chlorite		.47	.32
Limonite		2.23	1.50
Totals	99.09	100.00	67.50

[1] Includes other hydrous aluminum minerals. (See p. 309.)

On the assumption that alumina has remained constant (see pp. 8–9); orthoclase has suffered a loss of 6.95 grams or 41.9 per cent of the amount originally present. The fresh rock contained 53.1 per cent of plagioclase feldspars, which is reduced to 13.46 per cent in the altered rocks, representing a loss of 74.7 per cent of the plagioclase of the original rock. The percentage loss for the albite molecule is 70.4 per cent and for the anorthite molecule 84.5 per cent. These may be regarded as minimum losses, for, in calculating the mineral composition of the altered rock, all of the postah, soda, and lime, were calculated as feldspars. Since it is probable that part of these constituents, particularly the potash, is held by adsorption in the clay minerals, the percentage of feldspars calculated for the altered rock may be in excess of the amounts actually present.

The biotite of the fresh rock has been completely altered to chlorite, limonite, and kaolin, the loss of magnesia being 83 per cent of the amount present in the fresh rock.

Free quartz has been reduced by solution from 22.41 grams in 100 grams of fresh rock to 8.62 grams in the residual material, or a loss of 61.6 per cent of the original quartz. If there is less feldspar in the altered rocks than above calculated there is a higher percentage of quartz. If the primary silicates had altered completely to kaolin, the uncombined silica, figured as quartz, would be 26.4 per cent of the altered rock, or 17.8 grams, assuming alumina to have remained constant. This limiting case would represent a reduction from 22.41 grams of quartz in the fresh rock to 17.8 grams in the altered rock, or a loss of 20.6 per cent of the original quartz. It may be concluded then that there has been a loss by solution of between 20 and 60 per cent of the original quartz.

KATAMORPHISM OF GRANITE IN TERMS OF CHANGES IN CHEMICAL COMPOSITION

Georgia Granite as a Typical Case

Inspection of the analyses of Georgia granite (Table II) shows that the processes of decomposition have changed very markedly the *relative* amounts of the several constituents. The change in composition is due to the addition of certain components and the removal in solution of certain soluble substances.

TABLE II. ANALYSES OF FRESH AND ALTERED GRANITE FROM GEORGIA [1]

	I.	II.	III.	IV.
SiO_2	68.75	57.04	38.50	—30.25
Al_2O_3	17.59	26.04	17.59	0.00
Fe_2O_3	1.40	1.91	1.29	— .11
MgO	.64	.17	.11	— .53
CaO	3.25	.75	.51	— 2.74
Na_2O	4.54	1.91	1.29	— 3.25
K_2O	3.27	2.41	1.63	— 1.64
H_2O	.56	9.77	6.59	+ 6.03
	100.00	100.00	67.51	—32.49

I. Fresh granite from a depth of about 15 feet.
II. Decomposed granite from a depth of about 5 feet.
III. Decomposed granite expressed in grams per 100 grams of fresh rock assuming Al_2O_3 to have remained constant in amount during the alteration.
IV. Losses and gains in grams during alteration of 100 grams of granite, assuming Al_2O_3 constant.

The chemical analysis of the fresh rock shows the relative amounts of the constituents by weight. It may be considered as representing the number of grams of each constituent in 100 grams of rock. The same is true of the analysis of the altered rock. As the alteration of the granite involves both gain and loss of materials, it is evident that the residual material from the alteration of 100 grams of fresh granite is likely to have a different weight. For this reason the absolute losses and gains cannot be determined by direct comparison of the analyses. Before quantitative comparisons may be made of the amounts gained and lost, the analysis of the altered rock must be reduced to a proper basis of total weight, for comparison with that of the fresh granite. In all available pairs of analyses typical of the decomposition of igneous rocks under surface conditions there is a relative increase in alumina and iron in the residual material, indicating that these two are the least soluble of the constituents of the rock. Iron, in the form of ferric oxide, is known to be extremely insoluble under oxidizing conditions, but in the ferrous form or in the pres-

[1] Watson, Thomas L., op. cit., p. 312.

ence of reducing agencies it is more readily soluble. This fact makes iron an uncertain factor in calculation. Alumina, however, forms only the one oxide, and in the weathered rock it exists in clay which is known to be very resistant to solvents, either oxidizing or reducing, under ordinary surface conditions in the absence of sulphuric acid. The very small amount of alumina in solution in river water, and its practical absence in chemical sediments, in cements, and in veins and replacements deposited by cold solutions, confirm the fact of its comparative insolubility. It may be inferred, then, that alumina remains very nearly constant in amount during the decomposition of the granite.

From the analyses of Georgia granite it is seen that 100 grams of fresh rock contain 17.59 grams of alumina and 100 grams of the altered rock contain 26.04 grams of that oxide. If the absolute amount of alumina has remained constant during the alteration it follows that 100 grams of fresh rock have yielded in this particular alteration $\frac{17.59}{26.04} \times 100$, or 67.5 grams of residual material.

The composition of the altered phase of granite may now be expressed in terms of grams of altered rock resulting from the alteration of 100 grams of fresh rock. Since the analysis of the altered rock represents the numbers of grams of each constituent in 100 grams, it is necessary to take $\frac{67.5}{100}$ of each of the items in the altered rock analysis to reduce it to a total of 67.5 grams. The third column of the table (Table II) has been derived in this manner. The gains and losses for each constituent may then be ascertained by direct comparison of the fresh granite analysis and the recalculated analysis of the altered granite (columns I and III of Table II).

Water, oxygen, and carbon dioxide have been gained, and silica and soluble salts of the bases have been removed in solution, the net change being a loss of 32.49 grams from 100 grams of fresh granite. The total loss of material is 38.52 grams.

Silica shows a very marked loss, 100 grams of fresh rock losing 30.25 grams of this oxide, or 44.1 per cent of all the silica present in the fresh rock. While the percentage loss of silica is not so great as that of the alkalies and alkaline earths, the absolute loss

is greater than all other losses combined. The silica lost is mainly that liberated through the decomposition of the silicates. There is a more or less definite and constant ratio between the silica and bases removed in weathering of a granite, this ratio being that of bases to silica in the silicates.[1] The liberation of silica by the alteration of silicates has been checked experimentally. By treating orthoclase with water containing carbonic acid Headden [2] obtained a solution which, upon evaporation, yielded a residue carrying over 40 per cent silica.

The quartz of the rock is less soluble than the silicates but it is nevertheless dissolved to an important extent. In the case here described, the loss is between 20 and 60 per cent. Other evidence of solubility of quartz is discussed on pages 52–53.

The analyses express all of the iron as in the ferric state in both fresh and altered rocks, indicating a loss of .11 gram, or 7.8 per cent of the total ferric oxide. It is probable that the iron is largely ferrous in the fresh rock and ferric in the altered rock, as shown by the change in color during alteration. If both ferrous and ferric oxides were determined there would appear a very large percentage loss of ferrous oxide and a considerable gain of ferric oxide, the change being due to the oxidation and partial solution of the ferrous oxide. If it be assumed that all iron in the fresh granite was in the ferrous form, the amount of oxygen necessary to accomplish this oxidation is 0.15 gram.

Magnesia shows a loss of 0.53 gram out of an original .64 gram, or 82.8 per cent. Lime shows a total loss of 2.74 grams, or 84.4 per cent of the amount originally present. Soda has been lost to the extent of 3.25 grams, or 71.6 per cent of the amount of soda in the unaltered granite. Potassa has been lost to the extent of 1.64 grams, which is 50.2 per cent of the original amount.

Water has increased by 6.03 grams during the alteration of 100 grams of granite. This is practically absolute gain, as only .56 gram was present in the unaltered granite. This largely represents water chemically combined in the minerals of the altered granite rather than water mechanically held in the openings.

[1] Steidtmann, E., A graphic comparison of the alteration of rocks by weathering with their alteration by hot solutions: Econ. Geol., vol. 3, 1908, pp. 405–406.

[2] Headden, W. P., Significance of silicic acid in waters of mountain streams: Am. Jour. Sci., vol. 16, 1903, pp. 180–181.

Rate of change in katamorphism of Georgia granite. While quantitative data in this case are insufficient to determine relative rates of change for the various constituents, there are other data from weathering of Georgia granites which warrant the inference that the rates of change for the various constituents during weathering are not uniform. Watson [1] gives analyses of a fresh granite and two stages in its alteration, which show that there is a tendency for silica, lime and soda, to be lost more rapidly during the early stages of alteration than during the later stages. The same is true to a less extent of iron, magnesia and potash. If the percentage of constituents were plotted as ordinates and the degree of alteration as absissæ, the lines for silica, lime and soda would show a steep gradient in the early part of the alteration, growing flatter and nearly horizontal toward the end. This is partly explained for silica by the fact that the solution of silica at first results from the decomposition of the silicates which liberates the silica, and only the less soluble quartz is left to work on. The lime-bearing minerals are in general the minerals most easily broken down and consequently lime is one of the first things to be lost during weathering. The loss of lime also liberates a certain amount of silica. When the magnesium-bearing minerals are attacked by surface solutions there develop a number of hydrous magnesium silicates like chlorite, serpentine and talc, which are relatively stable at the surface. Consequently there is less tendency for magnesia to go into solution than for calcium, which is not required for these secondary minerals stable at the surface. Potassa, like magnesia, is used in certain secondary minerals, relatively stable under surface conditions, which usually form during the breaking down of the feldspar. White mica, and especially sericite, is the more common of these minerals. Loss of potassa is also retarded by the relative stability of orthoclase as compared with plagioclase. Still further it has been shown experimentally that the colloidal clay formed has a peculiar affinity for potash, and adsorbs and retains much of it.[2] Soda, on the other hand, like lime, is used to a less extent in these more stable second-

[1] Watson, Thomas L., Granites and gneisses of Georgia: Bull. 9–A, Geol. Survey of Georgia, 1902, pp. 312, 325–327.

[2] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, p. 127.

ary minerals, and consequently largely removed in solution when the sodium silicates break down.

Interpretation of chemical changes in katamorphism of Georgia granite by aid of "straight-line" diagram. The absolute loss of constituents during weathering makes it impossible to compare directly analyses of the fresh and altered material. By recalculating the altered rock on the assumption of the constancy of alumina during the alteration, we are able to compare the two analyses and determine the changes which have taken place. Another method which does not necessitate the assumption of a single constant, and which shows at once the relative behavior

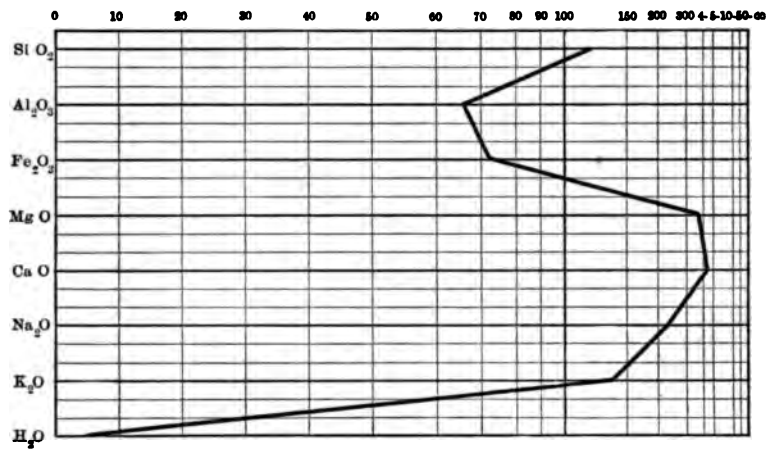


FIG. 1. Representing losses and gains of constituents in the katamorphism of a Georgia granite. (Table II, p. 8.)

of the several constituents during the alteration, is the "straight-line" diagram method, described on pages 288–289. In this the quotients obtained by dividing the percentage of each constituent in the fresh rock by its percentage in the altered rock, each multiplied by 100, are platted on a linear scale. The analyses of Georgia granite (Table II, p 8.) are represented in the form of a "straight-line" diagram in Fig. 1. This diagram is interpreted as showing that 67.5 grams of altered granite contain the same amount of alumina as 100 grams of fresh granite; that 120 grams of altered granite are required to contain the amount of silica

originally present in 100 grams of unaltered rock, and so on for the other constituents. It is obvious that alumina has increased relative to silica or that silica has decreased relative to alumina. The latter is more probable for reasons already stated. If alumina has remained constant, water has been increased and all other constituents lost during the alteration, the relative percentage lost for each constituent being indicated by the position of its point on the line.

Briefly then, if any constituent is assumed constant, all constituents whose points fall to the right of the assumed constant constituent have relatively decreased and those falling to the left have increased.

By this method it is not necessary to make new calculations to ascertain the relative gains and losses with reference to any substance assumed as constant. They can be read directly from the diagram. By inspection, the most probable points of constancy can be selected. For instance, in examining Fig. 1, it appears at once that soda could not be assumed to be constant, for this would involve a gain of all other constituents, resulting in a great increase in volume. On the other hand, alumina and iron remain nearly the same and are known to be relatively insoluble, hence it is the natural assumption that other substances have been lost relative to them.

Chemical Combinations of Substances Removed in Solution by Katamorphism of Granite

Alterations are above discussed in terms of gains and losses in residual materials. There is nothing in the analyses to show the chemical combinations in which the soluble constituents are removed and precipitated, but a study of the end-products and of the composition of ground, surface, and sea waters makes certain inferences possible. Silica is largely removed in the form of colloidal silicic acid [1] and also as alkaline silicates. It is probable that in minor amounts silica is also carried in solution as silicates of the other bases. Analyses of river waters and underground waters invariably show abundant carbon dioxide, and we may reasonably infer from this that the bases are to a very large extent carried as carbonates, particularly lime and magnesia. So

[1] Kahlenberg, Louis, and Lincoln, A. T., Solutions of silicates of the alkalies: Jour. Phys. Chem., vol. 2, 1898, pp. 77–90.

far as other acid radicals are present, such as sulphuric and hydrochloric acids, the bases are in part carried as sulphates and chlorides. Where pyrite is present in the rocks, its oxidation liberates sulphuric acid, which in turn combines with the alkalies and other bases, removing them as sulphates. This is particularly noticeable in the katamorphism of rocks associated with deposits of iron and copper sulphides.

These statements as to the combinations in which rock substances are carried in solution need qualification because of the fact that in dilute solutions the salts are supposed to be dissociated into their ions, and it is seldom possible to state definitely the manner of combination of the acid and basic radicals. Nevertheless the situation is approximately expressed by assuming the presence of salts of the sort above described in the solutions.

Summary of Chemical Processes of Katamorphism as Inferred From Chemical Changes in Granite

The products of weathering, both residual and transported, contain actual additions of the substances of the air and water, principally oxygen, carbon dioxide and water. These substances are the important agents in the alteration. Many other agents take part locally in the alterations to a large or even dominant extent, but are in general of minor importance as compared with the three principal agents already mentioned. Some of these agencies are organic acids, sulphuric and hydrochloric acids, alkaline sulphides and carbonates.

In the particular analyses here discussed (Table II) there are added in terms of grams, per 100 grams of fresh rock, 0.14 gram of oxygen, 6.03 grams of water, 2.74 grams of carbon dioxide, making a total of 8.91 grams added. In terms of elements there are added on the same basis 7.49 grams of oxygen, .67 gram of hydrogen, .75 gram of carbon, or, 84 per cent of oxygen, 7.52 per cent of hydrogen, 8.42 per cent of carbon.

Hydration, carbonation, and oxidation are therefore the important chemical processes of katamorphism. The water is principally in the clay, which is the most abundant product; in part also in the limonite and certain silicates. The carbon dioxide is in calcite and dolomite. The free oxygen added is in the limonite.

Chemical Changes in Katamorphism of Acid Igneous Rocks in General

The foregoing account of the alteration of granite will serve to bring out the essential features of katamorphism of acid rocks in general. Granites make up much the larger part of the acid igneous class within our zone of observation. Detailed consideration of all of the available data would not essentially modify the conclusions reached in the foregoing pages, but rather serves to verify and check them. In the quantitative study of katamorphism it is desirable to take advantage of all materials available, and with this in view, all of the alterations which have been chemically studied and the results of which are available to us, are assembled on a composite "straight-line" diagram.

In Plate II, 16 pairs of analyses representing weathering of acid igneous rocks are shown. Following is a list of references to the analyses platted.

1a. Granite gneiss, Georgia. Watson, T. L., Granites and gneisses of Georgia, p. 327.
1b. Biotite granite, Georgia. Watson, T. L., Granites and gneisses of Georgia, p. 321.
1c. Biotite granite, Georgia. Watson, T. L., Granites and gneisses of Georgia, p. 325.
1d. Biotite granite, Georgia. Watson, T. L., Granites and gneisses of Georgia, p. 318.
1e. Biotite granite, Georgia. Watson, T. L., Granites and gneisses of Georgia, p. 312.
1f. Feldspar, Georgia granite. Watson, T. L., Granites and gneisses of Georgia, p. 322.
1g. Granite gneiss, biotite, Georgia. Watson, T. L., Granites and gneisses of Georgia, p. 328.
1h. Biotite granite, Georgia. Watson, T. L., Granites and gneisses of Georgia, p. 309.
2. Gneiss, micaceous, Virginia. Merrill, G. P., Bull. Geol. Soc. of Am., vol. 8, 1897, p. 160.
3. Granite, District of Columbia. Merrill, G. P., Bull. Geol. Soc. of Am., vol. 6, 1895, p. 323.
4. Granite, Scotland. Holland, P., and Dickson, E., Proc. Liverpool Geol. Soc., vol. 7, pt. 1, p. 116.
5. Granite, Montana, ¼ mile southeast of Homestake. Analyses by C. T. Kirk.
6. Syenite, Arkansas. Williams, J. F., Ann. Rept. Geol. Survey of Arkansas, vol. 2, 1890, pp. 82 and 88.

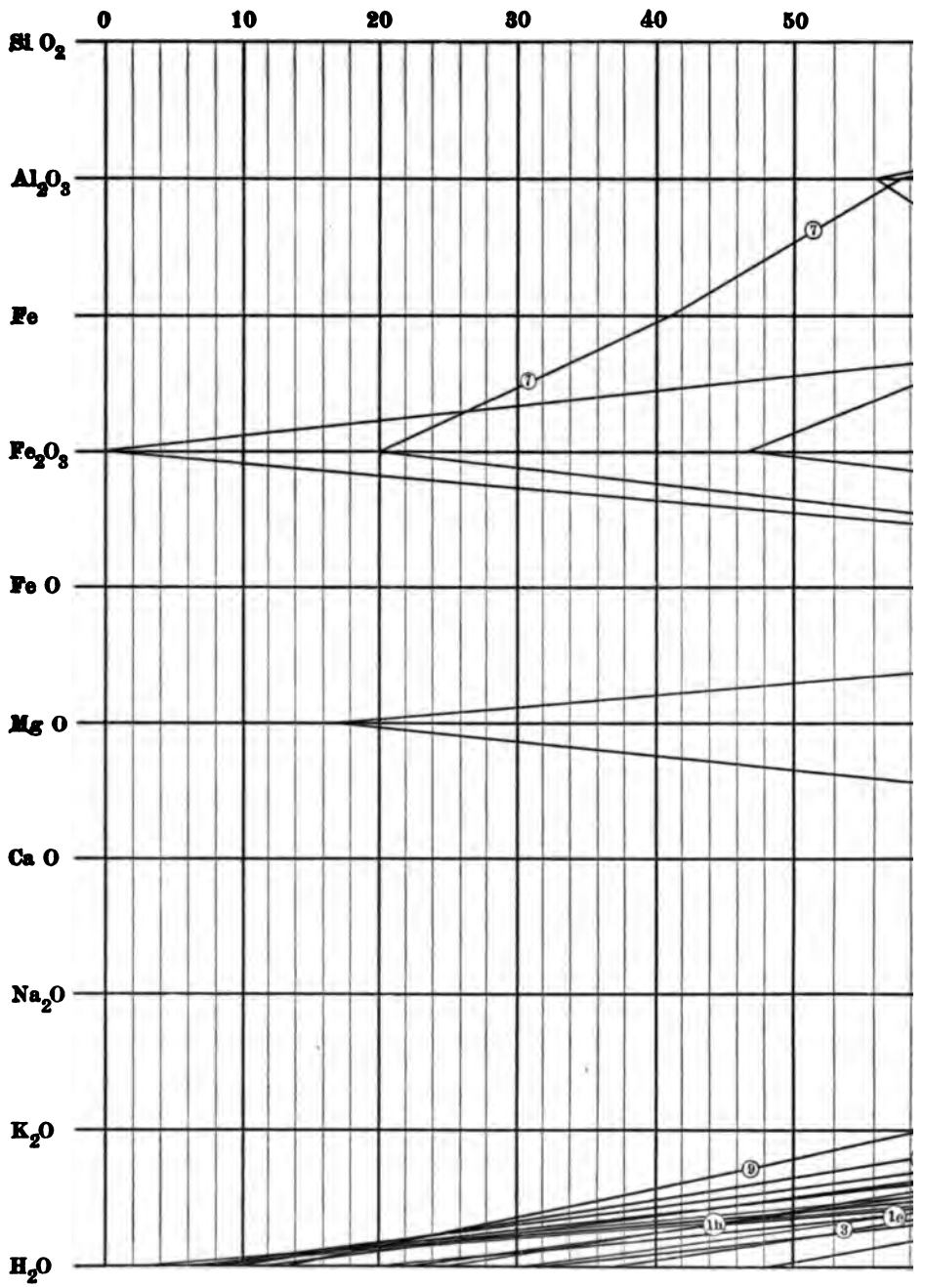
7. Andesite, Grenada. Merrill, G. P., Rocks, rock weathering, and soils: 1906, p. 208.
8. Quartz porphyry. Hähnel, O., Beitrag zur Frage der Kaolinbildung: Journ. f. Prakt. Chem., vol. 78, 1908, p. 281.
9. } Nepheline syenite, Arkansas. Mead, W. J., (see p. 30.)
9a. }

As in Fig. 1, a separate line is used for each constituent, and, to preserve the identity of the several analyses, the points for the various constituents in a given pair are connected and the connecting lines numbered to identify the analysis. These connecting lines serve a double purpose, in that they not only mark the points but they also indicate, by their degree and direction of slope, the behavior of a constituent relative to other constituents. For example, all of the lines between silica and alumina points slope downward to the left, indicating that in all cases silica has been lost relative to alumina. The lines with steep slopes show the smallest degree of change and those having flatter slopes, greater degrees of change. These lines serve to give a composite of the alterations platted.

This composite diagram shows that silica is in all cases lost relative to alumina. Iron in the majority of cases remains nearly as constant as alumina, but in some cases acts in a more variable manner. It may show increase or loss, depending on whether local conditions are oxidizing or reducing. Rocks altered by oxidizing solutions at or near the surface, tend to retain their iron. Rocks weathered by reducing solutions lose iron by removal in a soluble ferrous form. Illustrative of this is No. 8 (Pl. II), a quartz porphyry altered beneath a moor, where organic acids and the absence of oxygen cause the removal of iron. The oxidation and precipitation of these soluble salts may result in actual enrichment in iron. Case 7 (Pl. II) illustrates this. Losses or gains of ferrous and ferric iron are obviously a question of the presence or absence of oxygen. It will be noted that those cases which show a retention of total iron or an increase of total iron also show an increase in ferric oxide, the insoluble form of iron, and a corresponding decrease of ferrous oxide due to oxidation. Ferrous oxide is lost in all cases of oxidation and of solution of ferrous iron.

Magnesia in the majority of cases has been lost to a slightly

PLATE II



Representing losses and gains of constituents in weathering of acid i
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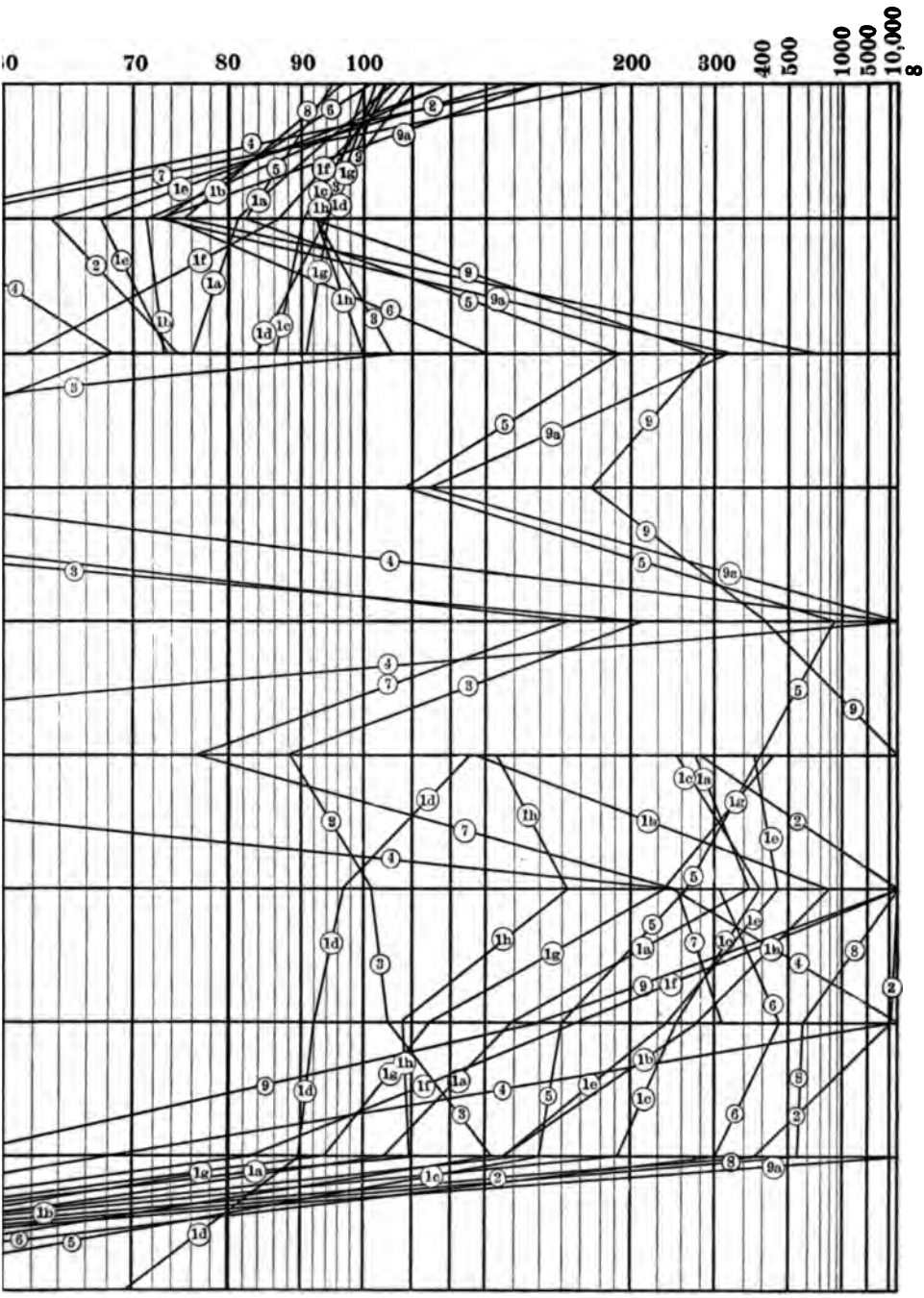


s rocks. Each broken line represents comparison of a pair of analyses, lentification.

greater extent than silica. Magnesia behaves in a rather variable fashion. In Nos. 3 and 7 (Pl. II) there is a distinct tendency for magnesia to be retained (in chlorite) about as much as alumina. In other cases, magnesia has been lost to a considerable extent.

Lime has been lost to a greater extent than magnesia, due to the retention of magnesia in some of the comparatively stable hydrous magnesium silicates.

Soda shows a smaller loss than lime and possibly a larger loss than magnesia. Potash is retained relative to soda and lime, much the same as magnesia and for similar reasons, i. e., the retention of potash as a constituent in the relatively stable minerals, orthoclase and sericite. It is retained also in somewhat indefinite solution in colloidal clay.

Water, as should be expected, has increased in amount in every case, due to hydration.

In all probability such variations from the common case as is shown by magnesia in No. 4 represent difference due to something other than normal weathering, possibly to samples not properly comparable.

VOLUME AND DENSITY CHANGES IN THE ALTERATION OF A GRANITE BY WEATHERING

General Consideration of Changes

Weathering of granite increases the volume: (a) by mechanically disintegrating and separating the mineral particles, forming openings, (b) by the addition of water, carbon dioxide and oxygen from the air, and (c) by decreasing the average density or specific gravity of the minerals and thereby increasing their volume. This summary statement compares all of the products of the weathering of a granite with the original granite. If the residual products of weathering alone are considered it is apparent that elimination of soluble constituents tends to reduce the volume and thereby to offset the effects of the above named factors. The net result may be either an increase or decrease of volume so far as the residual material is concerned. When both the residual material and that carried off in solution are considered the change is always an increase.

In the weathered residual, openings or pore space develop only

in proportion as material is removed by solution, and, therefore, involve no increase of volume, unless accompanied by a mechanical disturbance increasing porosity. Pore space in the residual material ranges from 1 per cent of the volume of the unaltered rock up to a maximum of 50 or 60 per cent with an average of about 35 per cent.

The increase in volume, due to hydration and the formation of soft minerals, causes the rock to fall to pieces near the surface, but beneath the surface the altered rock is relatively solid, perhaps because of the pressure exerted by the tendency to increase volume.

Material removed by solution is in part redeposited with a certain amount of pore space. In the case of a granite the material removed in solution is principally silica, soda, and potassa. The silica may be later deposited as chert, which is characteristically dense. Soda and potash salts are not reprecipitated to any extent as chemical sediments.

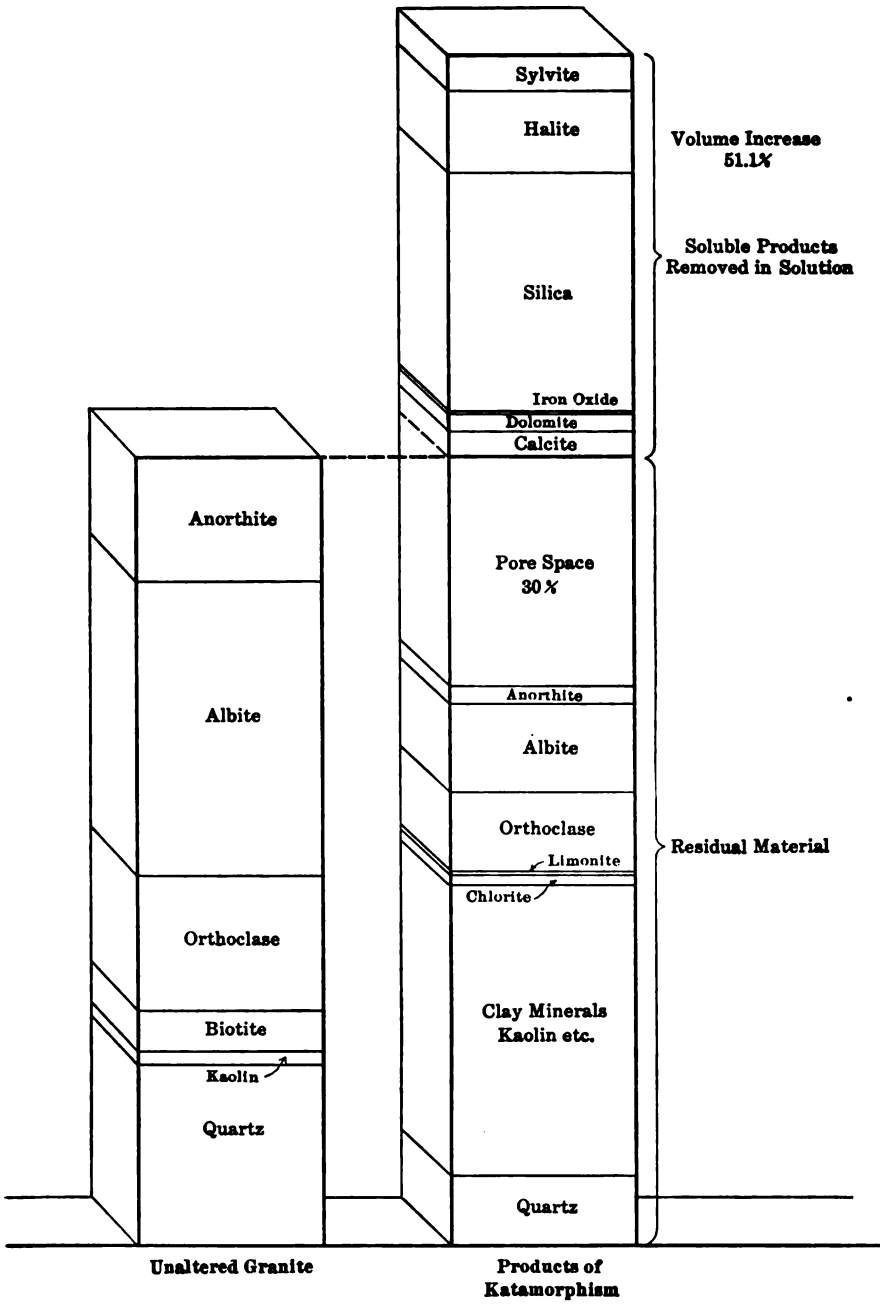
When the residual material of the granite becomes disintegrated, transported, and redeposited as fragmental sediments, principally mud and sand, the percentage of pore space becomes approximately 30 per cent for the unconsolidated mud and sand.

The specific gravity of an average granite (without taking into account pore space) is 2.68. On the assumption that the end-products of weathering are principally clay and quartz in the ratio of 75 to 25, the average specific gravity of the alteration product is 2.63 which represents an increase in volume due to mineral change of only 1.80 per cent. This is small as compared to the increase due to development of pore space in disintegration.

Consideration of Volume Changes in Alteration of Georgia Granite

The mineral composition of the fresh and altered Georgia granites given in Table I, page 6, has been expressed in terms of volume in Table XXIII, page 311. The volume compositions of the fresh rock and of the altered rock, on the basis of alumina constant, are represented graphically in Plate III. The volume of the residual material is 70 per cent of the volume of an equivalent amount of fresh rock, considering in both cases only the volume of the solid minerals. If the rock volume has remained constant during alteration, the difference in mineral volume, amounting to 30 per cent of the volume of the fresh rock, is rep-

PLATE III



Graphic representation of volume change in weathering of a Georgia granite.

resented by voids in the altered rocks. In other words, if no mechanical packing or slumping has taken place, the altered rock should have a porosity of 30 per cent of its volume, due to the leaching of soluble constituents.

By making certain assumptions as to minerals, the volume of materials removed in solution may be estimated. If all of the lime and magnesia removed in solution are figured as dolomite and calcite, the iron as limonite, the soda and potash as chloride, and silica as quartz, it appears that there are 47.79 grams of dissolved material per 100 grams of fresh granite. (See Table XXV, p. 313.) In terms of volume they are equivalent to 51.1 per cent of the volume of the original rocks. If it is assumed that unit volume of residual rock corresponds to the same volume of altered rock, in other words, that pore space is developed and is not closed by slumping, then this figure of 51.1 per cent expresses the increase in volume, due to katamorphism of the granite. These various considerations are expressed graphically in Plate III.

Part of the above increase in volume is due to actual addition of substances from the atmosphere and hydrosphere. To each 100 grams of fresh granite there is added by weathering .14 gram oxygen, 6.03 grams water and 2.74 grams carbon dioxide or a total of 8.91 grams. If the original rock and the products of alteration had the same average specific gravity the increase in volume due to addition of material would be 8.91 per cent. Since the products of alteration are of lower average specific gravity than the original rock the volume increase is greater than the above.

CHAPTER II

KATAMORPHISM OF BASIC IGNEOUS ROCKS

KATAMORPHISM OF BASIC IGNEOUS ROCKS IN TERMS OF ROCKS AND MINERALS

When basic igneous rocks are disintegrated without much chemical decomposition, the result is a dark green ferromagnesian sand. When chemical decomposition, transportation, and deposition take place, clays and iron oxides or iron carbonates, limestone and dolomite formations result. No pure sand formations are produced, as from acid igneous rocks, but there are correspondingly greater amounts of limestone, dolomite, iron oxide and iron carbonate. The ultimate mineral products are clay, calcite, dolomite, iron oxide, and iron carbonate. As intermediate steps in the alteration of the basic feldspars to their end-products, there may be developed epidote, sericite, and more acid feldspars, in all cases with liberation of lime. From the augite there are developed hornblende and chlorite, with separation of calcite. Not uncommonly the hornblende is of a fibrous or uralitic variety. Chlorite develops from the hornblende. Serpentine or talc may develop either from the augite or hornblende. Bastite develops from augite. Biotite alters to chlorite and magnetite. Ilmenite changes to leucoxene. Olivine alters to serpentine and magnetite. It alters more readily than other minerals, allowing free circulation of water through the rock and thus promoting the alteration of other minerals.

KATAMORPHISM OF BASIC IGNEOUS ROCKS IN TERMS OF CHEMICAL COMPOSITION

When a basic igneous rock, such as a gabbro or diabase, is exposed to the decomposing action of air and water it behaves very much like the granite. The silicate minerals are broken down to form oxides and carbonates, and silica is liberated. As in the case of the granite, some of the new substances formed,

particularly the carbonates, are soluble and are removed in solution, the residual material consisting of insoluble products such as clay and iron oxide, together with the undecomposed minerals. There are, however, certain general differences between the behavior of the basic igneous rock and granite, due principally to their difference in composition. The absence of free quartz in basic rocks, the smaller proportions of combined silica, and the correspondingly larger proportions of the bases, makes the basic rock much more susceptible to attack by the agents of weathering, which are mainly acid in their nature. More energy is liberated by weathering of such rocks than of acid rocks. The weathering of basic rocks, therefore, proceeds more rapidly than that of acid

TABLE III. ANALYSES OF WEATHERED BASIC IGNEOUS ROCKS

	Weathering of basalt [1]			Weathering of diabase [2]		
	Fresh	Altered	$\frac{\text{Fresh}}{\text{Altered}} \times 100$ [3]	Fresh	Altered	$\frac{\text{Fresh}}{\text{Altered}} \times 100$ [3]
SiO_2	45.21	59.98	75.3	43.56	44.93	97.
Al_2O_3	7.82	11.50	68.	14.58	16.27	89.6
Fe_2O_3	3.41	2.42	141.	3.84	13.37	28.7
FeO	8.08	1.86	435.	7.	0.0	∞
MgO	8.43	.75	1123.	9.95	6.40	155.3
CaO	12.31	2.80	440.	10.78	1.84	586.
Na_2O	6.64	1.18	563.	1.86	2.03	91.7
K_2O	2.94	1.48	199.	1.02	.84	121.5
H_2O-						
H_2O+	1.82	16.53	11.	3.85	12.55	30.7
SO_3	.56	.21				
Fe	8.69	3.14	276.	8.15	9.37	87.0
TiO_2				1.03	1.34	77.
CO_2				1.93		
Mn				.39	.28	

[1] Endell, K., Über die chemische und mineralogische Veränderung basischer Eruptivgesteine bei der Zersetzung unter Mooren: Neues Jahrbuch für Miner., Geol. und Paleo., Beilage-band 31, 1911, p. 18.

[2] Holland, P., and Dickson, E., Remarks on the formation of clay: Proc. Liverpool Geol. Soc., vol. 7, 1892–1893, p. 108.

[3] These columns show for each constituent the number of grains of altered rock required to contain the amount of that constituent originally present in 100 grams of fresh rock.

rocks. When the two are in association, the weathered basic rocks often are in topographic depressions, while the more resistant acid rocks form the ridges. In the preceding table are analyses showing two cases of weathering of basic igneous rocks. The first is a case of weathering of a basalt beneath a bog and the second is a case of surface weathering of a diabase. In the first case the rock has been altered under reducing conditions with little or no opportunity for oxidation. In the second case there is abundant opportunity for oxidation, as the rock is exposed at the surface. These alterations are both shown graphically in Fig. 2 by means of the "straight-line" diagram. Both analyses

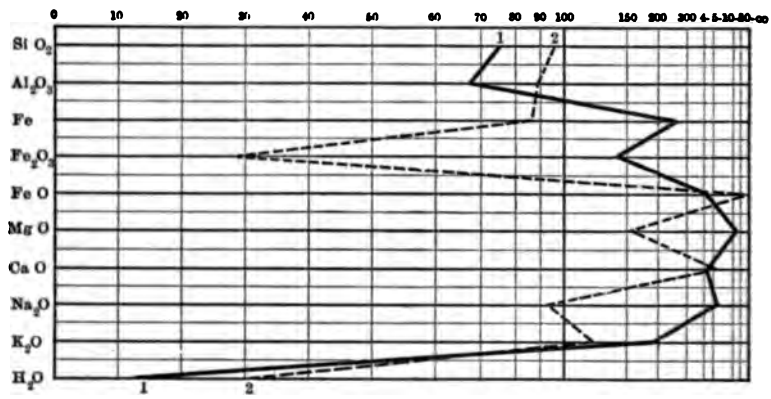


FIG. 2. Representing the losses and gains of constituents in the katamorphism of basalt (solid line) and diabase (dash line) represented by analyses on page 21.

show silica lost relative to alumina. Iron in the first case has decreased relative to alumina and in the second case has increased. Magnesia, lime, soda and potash show varying degrees of loss relative to alumina. Water has been increased in both cases.

No marked difference between the two sets of analyses is observed except in the behavior of the iron. This is the characteristic difference found between rocks weathering at the surface under oxidizing conditions and those weathering below the surface where oxygen is absent or less abundant in the solutions. Iron, because of the insolubility of the ferric oxide, is extremely insoluble in oxidizing solutions, but in the ferrous form, as car-

bonate or sulphate, or salts of organic acids, it is comparatively soluble, and hence is easily dissolved and transported by reducing solutions. If these reducing solutions carrying iron travel to the surface, the iron is precipitated as ferric oxide as soon as these solutions mingle with oxidizing solutions, or acquire oxygen from the air, or meet iron bacteria. This process may result in an actual increase of iron in the oxide zone even to the extent of producing a bog iron ore. It is commonly the case that rocks actually acquire iron in the zone of oxidation and lose iron when they are altered below the zone of oxidation.

Plate IV is a composite of 17 cases of weathering of basic igneous rocks represented in the same manner as the composite for acid igneous rocks in Plate II. Following is a list of references to the analyses platted.

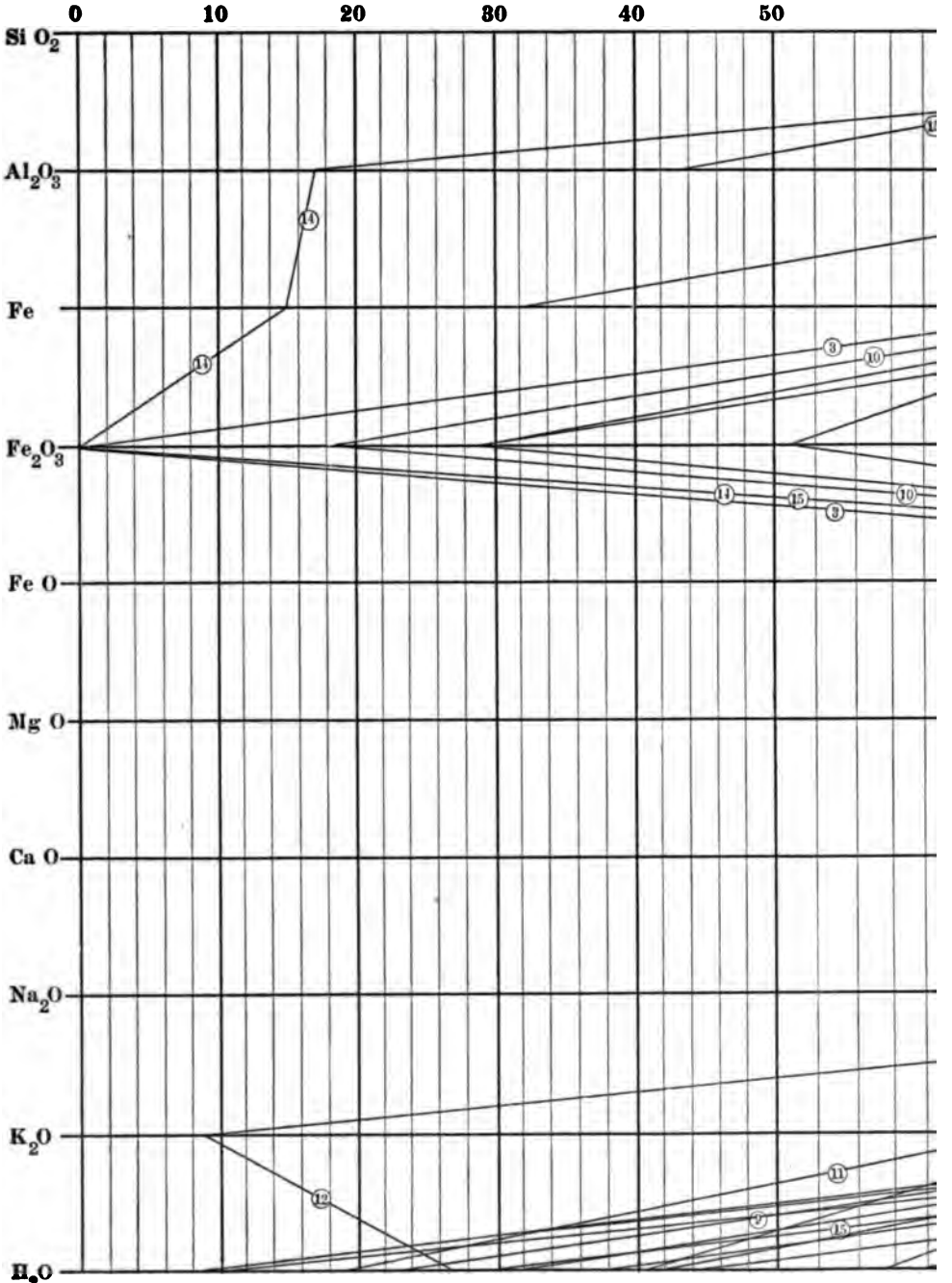
1. Diabase, Island of Jersey. Holland, P., and Dickson, E., Proc. Liverpool Geol. Soc., vol. 7, 1892–93, p. 108.
2. Diabase, Massachusetts. Merrill, G. P., Disintegration and decomposition of diabase at Medford, Massachusetts: Bull. Geol. Soc. Am., vol. 7, 1896, p. 353.
3. Diabase, Spanish Guiana, Venezuela. Clarke, F. W., Bull. 330, U. S. Geol. Survey, p. 413.
4. Diorite, Virginia. Merrill, G. P., Rocks, rock weathering and soils, 1906, p. 225.
5. Phonolite, Bohemia. Merrill, G. P., Rocks, rock weathering and soils, 1906, p. 198.
6. Gabbro, Minnesota. Boulder near Allen Junction.
7. Diabase, Chatham, Virginia. Watson, T. L., Weathering of diabase near Chatham, Virginia: Am. Geologist, vol. 22, 1898, p. 87.
8. Basalt, Bohemia. Merrill, G. P., Rocks, rock weathering and soils, 1906, p. 223.
9. Trap rock, Connecticut Valley. Hawes, G. W., Am. Jour. Sci., No. 51, vol. 9, March, 1875, p. 185.
10. Dolerite. Rowley Regis, England.
11. Augite andesite, Washington.
12. Diabase (dike), Penokee-Gogebic District, Mich.
13. Augite diorite, Magnetberg, Southern Urals. Clarke, F. W., Bull. 330, U. S. Geol. Survey, p. 413.
14. Serpentine rock, Cuba. Unpublished analyses.
15. Basalt, Haute Loire, France. Merrill, G. P., Rocks, rock weathering and soils, 1906, p. 206.
16. Alnoite, Herkimer County, N. Y. Smyth, C. H., Jr., Bull. Geol. Soc. Am., vol. 9, 1898, p. 262.
17. Basalt, Bauersberg. Endell, K., Neues Jahrbuch, B. 31, 1911, p. 18.

An inspection of the composite diagram shows that silica has been lost relative to alumina. The total loss is less than in acid igneous rocks, for the reason that the breaking up of the basic minerals frees less silica. Iron in some cases has been lost and in other cases has increased relative to alumina, depending, in all probability, on the presence or absence of oxygen in the weathering solutions. Magnesia and lime have been lost, but there appears to be a tendency for the lime to be lost to a greater extent than the magnesia. Both soda and potash show a loss. Soda is lost more than potash. Water has been increased in all cases.

VOLUME CHANGES IN KATAMORPHISM OF BASIC IGNEOUS ROCKS

The same factors enter into the problem as in the weathering of an acid igneous rock, and will not here be discussed in detail. Taking into account both residual and transported minerals, there is increase of volume through disintegration and development of pore space, there is an increase of volume by the addition of water, carbon dixoide and oxygen from the air, there is an increase of volume by the decrease of the average density or specific gravity of the minerals. The most important factor is the development of pore space, which may be calculated in a typical case to be about 12 per cent. Increase of volume, due to change in density of the minerals, amounts to about 11 per cent. It will be noted that while the development of pore space is about the same in the basic and acid igneous rocks, the increase of volume, due to change in mineral density, is much greater for the basic igneous rock.

PLATE IV



Representing losses and gains of constituents by weathering of basic ign
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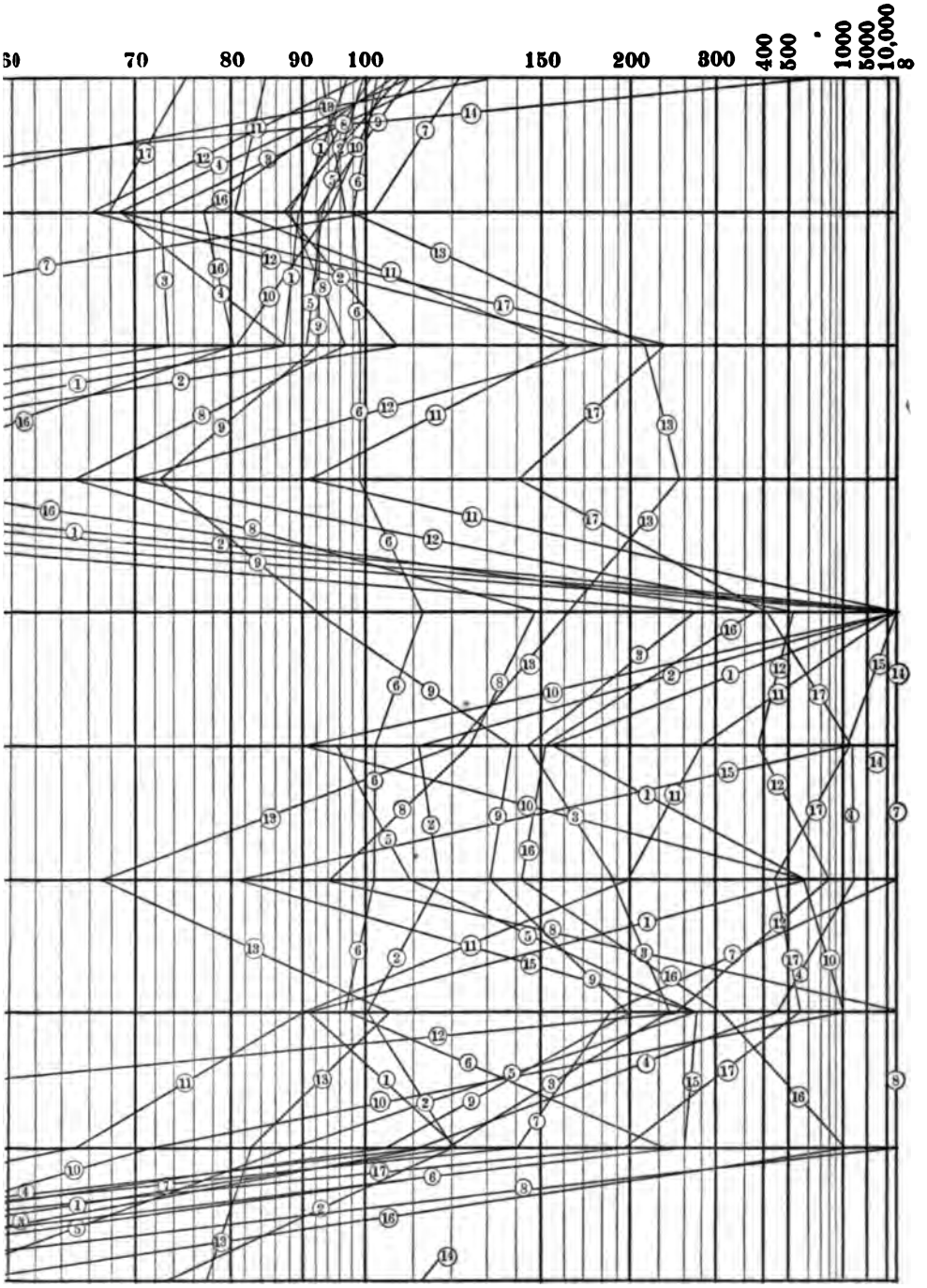
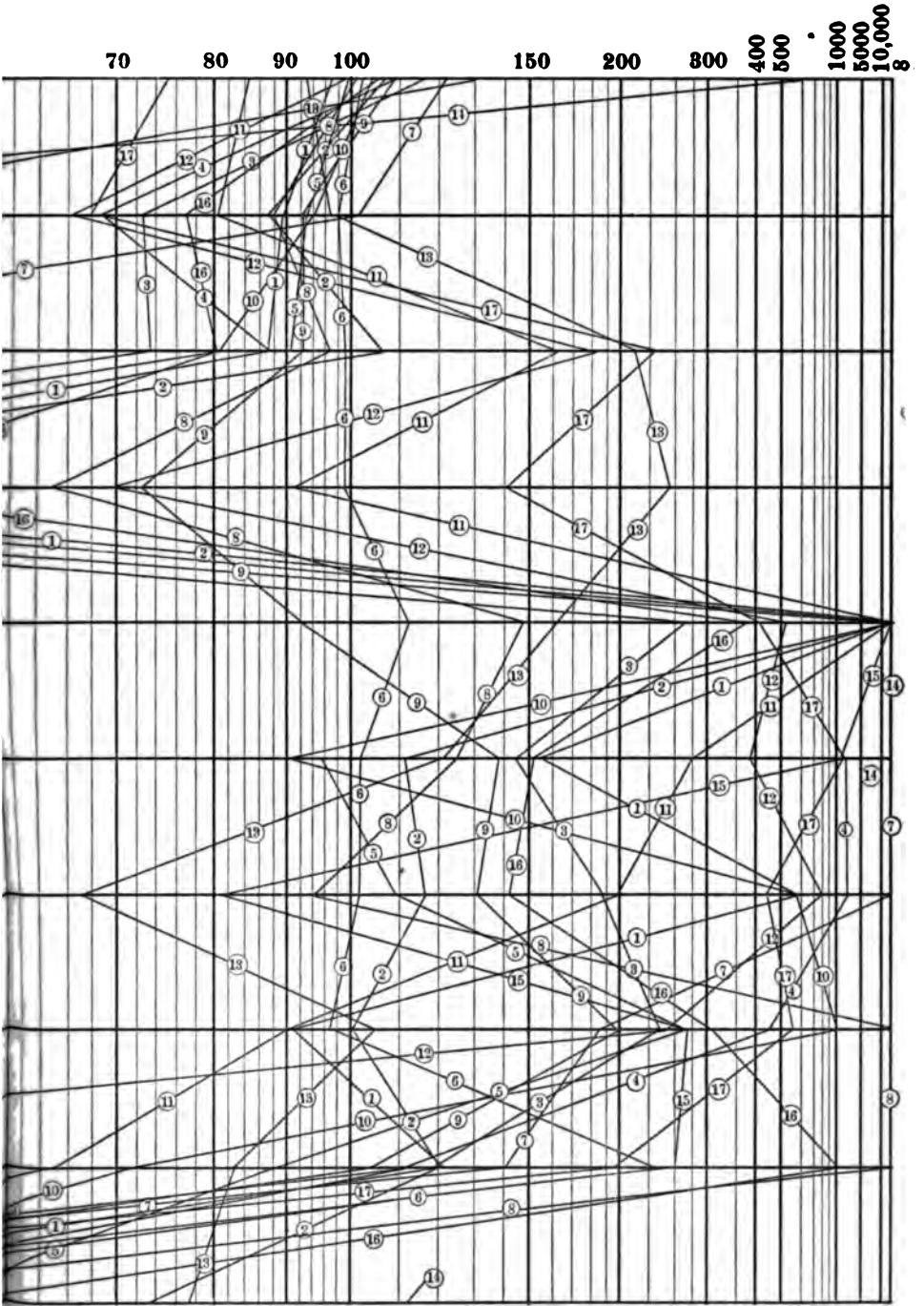


us rocks. Each broken line represents comparison of a pair of analyses, dentification.



rocks. Each broken line represents comparison of a pair of analyses,
tification.

of a nepheline syenite. Some of the bauxite is interstratified with Tertiary sediments and has clearly been transported.

Color and Texture

The bauxite varies greatly in color and texture. The prevailing color is light buff, which grades into white, and various shades of yellow, brown, and red, depending on the content of ferric iron. It varies in hardness from that of a limestone or well cemented sandstone, to a soft material which can be shoveled without picking. The hard portions of the bauxite constitute the upper part of the deposit and the soft portions are found below. The upper portion of the ore has suffered a marked induration or cementing process. The bauxite has a comparatively high porosity, an average of a number of tests showing 38.5 per cent pore space. The free moisture varies greatly with the season of the year. The moisture content (uncombined water) varies from 5 per cent to a maximum of 20 per cent, with an average between 12 per cent and 15 per cent.

The prevailing texture is pisolitic or "oolitic" common to bauxite, made up of more or less spherical pisolites of bauxite, sometimes having concentric structure, in a bauxite matrix. These vary in size from microscopic to a maximum of an inch or more in diameter. Second in importance is the ore having the texture of the syenite, commonly called "sponge ore" or "granitic ore." This bauxite preserves in varying degrees the original granitic texture of the syenite. The granitic type of ore is generally found beneath the oolitic ore and grading upward into it. Granitic bauxite grades downward into kaolinized syenite, and this in turn into unaltered syenite. Some of the bauxite has an amorphous texture having the appearance of clay, from which it is distinguished with difficulty.

Chemical Composition

The principal chemical constituents of the ore in order of their abundance are alumina, combined water, silica, ferric oxide, titanium oxide, ferrous oxide, and carbon dioxide. In every deposit of bauxite, excluding the minor constituents which remain essentially constant, there is a complete gradation from the com-

position of gibbsite (Al_2O_3=65.5 per cent, H_2O=34.5 per cent) to the composition of kaolin (Al_2O_3=39.5 per cent, SiO_2=46.5 per cent, H_2O=14.0 per cent. (See Fig. 5.)

In addition to the constituents above mentioned, the bauxite contains minute amounts of zirconia, barium, iron sulphide, and other accessory constituents.

The triangular diagram (Fig. 5), illustrates graphically the variations in chemical composition of the ores.

Mineralogical Composition

The principal mineral constituents of the ore are the two forms of hydrous aluminum oxide,—the crystalline form *gibbsite*, and the amorphous form *bauxite*, both approximating the formula $Al_2O_3.3H_2O$. It should not be inferred from this that gibbsite exists in commercial quantities as such. The two forms of the trihydrate of aluminum differ only in that one is crystalline and the other amorphous and neither can be mined separately. The granitic type of ore is largely composed of microcrystalline aluminum trihydrate. The pisolitic type of ore is made up largely of the amorphous form, but this type of ore also contains a considerable amount of the crystalline form in the shape of remnants of feldspars which have been completely altered to gibbsite. (Fig. 3.) It would seem from this that the hydrous aluminum oxide develops first in the form of gibbsite and that when it undergoes change, as in the development of the pisolitic type of ore, it tends to take on the amorphous form.

Hydrous aluminum silicates, of which the most abundant is kaolinite, are second in importance. Halloysite, or a closely similar compound, is known to be abundant both by recalculation of chemical analyses and its observed presence in the lean portion of the bauxite. In addition to these, there are in all probability other hydrous aluminum silicate minerals which are difficult to identify because of their amorphous texture. Limonite and other hydrous oxides of iron are undoubtedly present in the ores, and it is also probable that some of the ferric iron is combined with the alumina in the hydrous aluminum oxides.

Siderite occurs in a very irregular manner in the ores. It is very evident that it has been introduced subsequent to the forma-

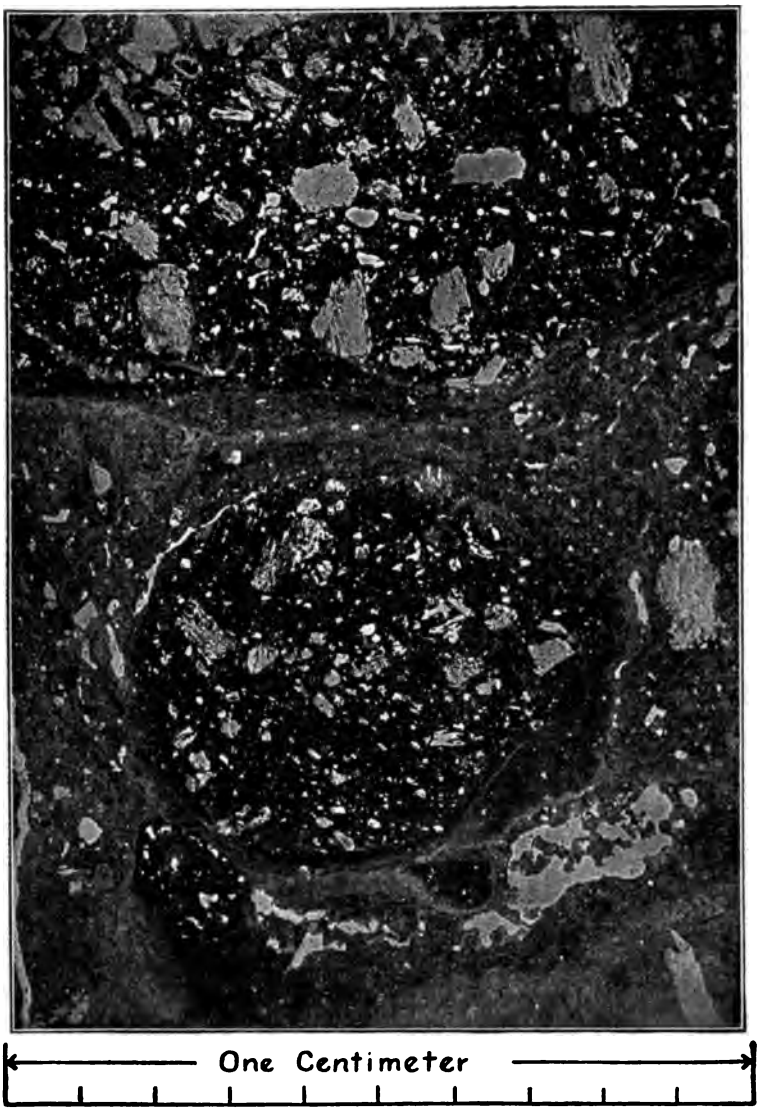


FIG. 3. Photomicrograph of a thin section of pisolitic bauxite from Arkansas. Ordinary light. (One complete pisolite and a portion of a larger one are shown). The darker portions are amorphous bauxite stained brown and red by iron oxide. The irregular white area in the lower right-hand corner is a hole in the section. The roughly rectangular light colored areas are feldspar forms now microcrystalline gibbsite. The form and structure of the feldspars are preserved and in polarized light effects of feldspar twinning may be seen.

tion of the bauxite. Titanium is present as titanite and ilmenite, and also in all probability in the form of other titanium minerals

In addition to the above minerals, a very interesting group of heavy accessory minerals is found in small quantity in the bauxite and associated clays. They are easily concentrated from the bauxite, because of their high specific gravity, by crushing and panning the ore and separating by means of a heavy solution. In an incomplete examination of these minerals the following were identified: zircon, occurring in a variety of crystal forms and varying in color from white to dark brown through shades of yellow; pyrite; barite; celsian; and fluorite. There are other minerals which were not identified. These heavy accessories are excellent evidence of residual nature of the bauxite and associated clay, since the same group of heavy minerals is also found in the unaltered syenite beneath the bauxite.

Origin of the Bauxite Deposits

A study of the general geology, chemistry, and mineralogy of the deposits has led to the following conclusions:

(1) The bauxite and associated clays are the products of surface weathering of the syenite by normal processes of rock decomposition, and are in no sense chemical sediments.

Bauxite deposits occurring on the syenite surface have developed in situ from the syenite.

The deposits developed in situ from the syenite show evidence of downward secondary concentration of alumina.

Bauxite lenses occurring interstratified with the Tertiary sediments consist of material which has been removed from its place of origin by Tertiary streams.

(2) The texture of the kaolinized syenite has been essential to the alteration of the kaolin to bauxite.

(3) The oolitic or pisolitic texture of the bauxite has developed in place from the granitic or amorphous types of bauxite.

Evidence that the deposits were produced by weathering of the syenite. (1) That there is some genetic relationship between the syenite and the bauxite is obvious from their distribution and has been recognized since the discovery of the deposits.

(2) The gradation from syenite through kaolinized syenite to bauxite is to be observed wherever a fresh cross-section is

available. This gradation, chemical, mineralogical, and textural, is at once suggestive of the residual nature of the deposits. The granitic texture of the syenite is preserved in the kaolinized syenite and in the granitic type of bauxite which grades into the pisolitic ore in which feldspar shapes replaced by gibbsite may be seen. The gradation from bauxite through kaolinized syenite occurs not only downward to the syenite beneath, but laterally and upward to residual boulders of decomposition of syenite within the bauxite deposit. These boulders, from a foot to twenty feet and more in diameter, are found in all stages of decomposition, some having cores of fresh syenite and others having been completely altered to bauxite and now represented only by boulder forms in the bauxite. They indicate in excellent fashion the residual nature of the deposits.

Chemical analyses of samples show a complete and uniform gradation in composition from the unaltered syenite through the kaolinized phase to the bauxite. There is no sharp break at any point in the gradation. The gradation from syenite to kaolin is shown, by numerous analyses, to be of a normal type of surface weathering. In the following table are analyses of samples from the same locality. No. 1 is fresh syenite from the cut near the Lantz mine, No. 2 is partially kaolinized syenite, No. 3 is completely kaolinized syenite containing some bauxite, and No. 4 is bauxite.

TABLE IV. ANALYSES OF A SERIES OF SAMPLES FROM THE SAME LOCALITY SHOWING GRADATION FROM UNALTERED SYENITE NO. 1 TO BAUXITE ORE NO. 4

	No. 1	No. 2	No. 3	No. 4
SiO_2	58.00	52.64	39.80	10.64
Al_2O_3	27.10	29.56	37.74	57.48
Fe_2O_3	1.86	1.06	1.60	2.56
FeO	3.30	.80	.10	.20
MgO	.25	.00	.00	
CaO	1.62	.00	.00	
Na_2O	6.70	4.46		
K_2O	.25	.44		
TiO_2	.40	1.20	3.30	1.20
H_2O	1.22	9.00	17.00	28.36

In Fig. 4 the volume compositions of the four samples is shown in terms of minerals and pore space. The porosity in each case was actually determined and the relative amounts of minerals calculated from the above analyses. The gradation is obvious.

In the triangular diagram (Fig. 5) a number of analyses of syenite, kaolinized syenite, clay, and bauxite, have been platted on the basis of chemical composition. Each of the small triangles

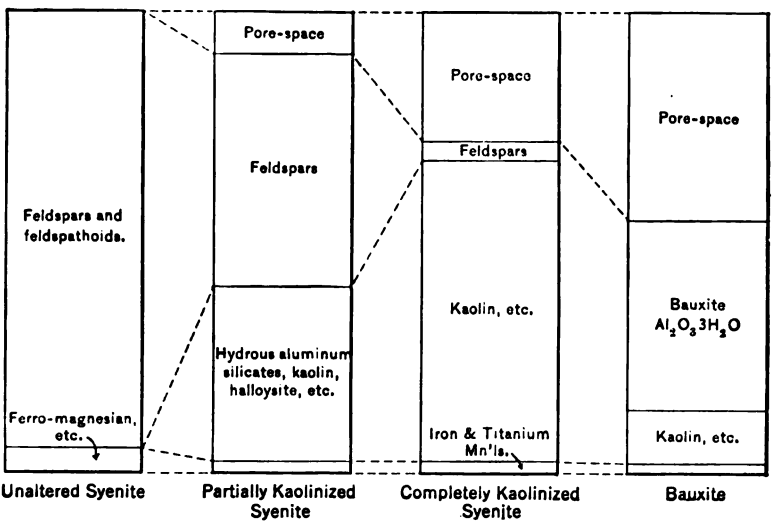


FIG. 4. Diagram showing gradation from syenite to bauxite in terms of volume. The columns represent a series of samples from a single locality.

represents a single analysis and by its position in the large triangle indicates the percentage of Al_2O_3, H_2O, and SiO_2. The distance up from the base of the large triangle indicates the percentage of Al_2O_3, the distance from the left side of the large triangle shows the percentage of H_2O and from the right side the percentage of SiO_2. The amount of Fe_2O_3 in each analysis is shown by the shaded area and may be read at the base of the small triangle. The gradation from the syenite through the kaolinized phase to the low grade bauxite and to merchantable ore is well shown here. It appears that normal weathering of the syenite to kaolin and allied minerals has simply continued one step further in breaking

down these hydrous aluminum silicates to hydrous aluminum oxide.

(3) The irregular and pitted surface exposed in the bottom of the open pit mines on the removal of the bauxite is in the nature of an etched surface, typical of decomposition, and is certainly not the type of erosion surface to be expected beneath a sedimentary deposit.

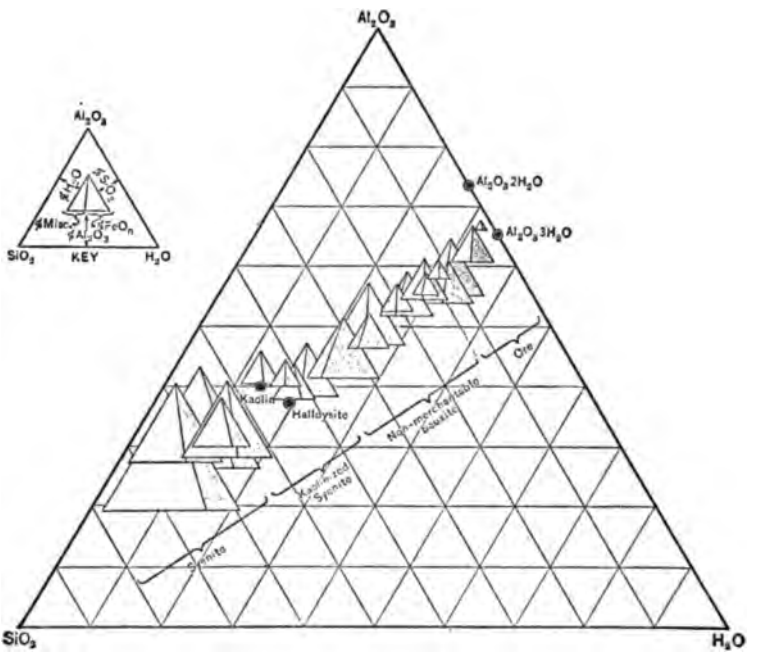


FIG. 5. Diagram showing gradation from syenite to bauxite in terms of principal chemical constituents. Each triangle on the diagram represents an analyzed sample.

(4) The bauxite beds are found through a vertical range of over 200 feet in the bauxite district alone, which fact, together with the uniform nature and thickness of the deposits, can hardly be reconciled with the idea which has been argued that the deposits are chemical sediments precipitated from solution in a shallow sea.

(5) The high porosity of the bauxite is taken as evidence of its

residual origin. The average porosity of bauxite is 38.5 per cent, and locally it is as high as 60 per cent, which is higher than would be expected in a sedimentary deposit either of chemical or fragmental origin. This porosity is in accord with the volume of material leached from the syenite in the formation of the bauxite.

(6) The bauxite and associated clays contain certain accessory minerals native to the syenite which have remained unaltered during the decomposition of the parent rock. These minerals are evenly distributed through the bauxite in a manner not possible were the bauxite of sedimentary origin. In the transported detrital portions of some of the bauxite, they are found more or less segregated into bands and bunches, as is to be expected, since the minerals are all of high specific gravity as compared with the bauxite and would tend to be concentrated by the action of water. That the bauxite found interstratified with the Tertiary sediments originated on the syenite surface is evidenced by the fact that abundant samples from drill holes yielded the same group of heavy accessory minerals as did the bauxite occurring in place on the syenite.

(7) That the changes above described are brought about by normal weathering processes and not by thermal waters, like hot springs, as supposed by Hayes,[1] seems clear from the uniform distribution of the different products of alteration in relation to the erosion surface, from the fact that the alterations are closely related to and grade into those of ordinary weathering, from the fact that they differ in distinctive features from the known results of alteration by thermal solutions, and finally from the widespread distribution of lateritic alterations of this type, through tropical countries, indicating certain climatic control, and not the local effects of more localized thermal solutions.

Downward secondary enrichment of alumina. The top portion of the bauxite deposits is characteristically higher in silica than the bauxite below. In many instances it has been necessary to remove the upper eighteen inches or two feet of high silica ore before mining. This increase in silica toward the surface may be due to an actual downward secondary concentration of alumina. Those portions of kaolin and halloysite which persist after the

[1] Hayes, C. W., The Arkansas bauxite deposits: 21st Ann. Rept., U. S. Geol. Survey, pt. 3, 1901, pp. 435–472.

main portion of the kaolinized syenite has been altered to bauxite are the dense impervious parts. Bauxite is soluble in surface solutions to a certain extent and on being dissolved and carried downward leaves the kaolin, and hence the silica, in relatively higher percentage at the surface. The alumina carried down is deposited below, assisting in the formation of the pisolites.

Open texture essential to change from kaolin to bauxite. The clay masses associated with the bauxite are of two types. One is a porous cellular phase preserving perfectly the texture of the syenite and consisting of a sponge-like aggregate of kaolinized feldspar individuals. It generally occurs in the lower portions of the deposits and is locally called kaolinized syenite or "underlay." This type is a transitional stage between syenite and bauxite and grades into both. The open texture is due to the openings left by the decomposition and removal of the easily altered mineral nepheline. The other type of clay is of a dense impervious nature, frequently occurring in large masses extending to the surface of the bauxite. Abundant slickensided partings occur, and color bands, marking former joint planes or the transient outlines of boulders of decomposition, are contorted and faulted. These facts are taken as evidence of slumping or packing of the kaolinized rock. There is no evidence whatever to indicate that the parent rock of the dense clay differed from the type of rock from which the porous clay was derived.

The occurrence of masses of unaltered clay at the surface of the bauxite raises the question as to why the clay escaped alteration to bauxite. An obvious explanation is that the impervious texture of the clay prevented circulation of solutions and made alteration to bauxite impossible, and that alteration to bauxite has proceeded only in those portions of the kaolinized syenite which have escaped this slumping or packing process. This suggests that the alteration of the nepheline syenite to bauxite has been possible because of the porous or spongy texture imparted to the kaolinized rock by the removal of the mineral nepheline.

The alteration of clay to bauxite, like other phases of rock decomposition, is accomplished through the agency of circulating solutions. It therefore seems reasonable to assume that the rate of alteration is dependent at least in part on the permeability of the clay. This suggests that the characteristic stability of clay

in general under surface conditions may be due to its equally characteristic imperviousness, and that only open textured clays may be altered to bauxite.

The formation of aluminum hydrates as residual products of rock weathering has been frequently observed and described. A summary of the literature on this type of alteration is given by Clarke in The Data of Geochemistry.[1] The formation of bauxite as a residual product of weathering has been observed for the most part in tropical and sub-tropical regions. It has been variously ascribed to the more active and complete leaching under tropical conditions of temperature and vegetation, to the action of bacilli or other microörganisms peculiar to the tropics and to nitric acid produced by tropical thunderstorms. Lateritization has been so generally correlated with tropical conditions that it seems probable that the Arkansas bauxite was developed under a tropical or sub-tropical climate.

The reason for the confinement of lateritization to tropical or sub-tropical regions has long been an open question. If, as has been suggested above, an open texture in clay or kaolinized rock is essential to the development of bauxite, the question arises as to why open textured clays should be peculiar to the tropics. A possible explanation to this may lie in the fact that the disintegrating action of frost on an open textured clay would be to destroy the cellular structure and allow the clay to slump and become impervious. This would tend to retard further alteration of clays in regions subject to frost and to confine lateritization to the regions of little or no frost.

Origin of the pisolitic or oölitic structure. This peculiar structure which appears to be common to bauxite the world over, has been adequately described by Branner [2] and Hayes [3] for the Arkansas deposits. The oölitic type of ore consists of an aggregation of pisolites or oölites varying in size from microscopic to a maximum of an inch, or in exceptional cases possibly more, in diameter in a matrix of bauxite. The pisolites frequently, but not always, have

[1] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, pp. 469–477.

[2] Branner, J. C., The bauxite deposits of Arkansas: Jour. of Geol., vol. 5, 1897, pp. 263–289.

[3] Hayes, C. W., The Arkansas bauxite deposits: 21st Ann. Rept., U. S. Geol. Survey, pt. 3, 1901, pp. 435–472.

a concentric structure marked by a variation of color bands, the coloring being due to ferric oxide. Occasionally oölites are observed to have a radiating system of cracks crossing the concentric banding. These cracks are usually cemented or filled by microcrystalline gibbsite. Some of the pisolites are compound, being made up of an aggregate of smaller individuals. Some of the pisolites or oölites are hollow at the centers and many of them have an extremely porous or spongy center within a comparatively thin shell of gibbsite. Many of the oölites, on being broken open, exhibit the texture of granitic ore, the feldspar shapes being very well and plainly preserved.

This peculiar pisolitic or oolitic texture, because of the fact that oölites have been thought to be the result of precipitation from solution in agitated waters, has in part led to the belief that the bauxite deposits themselves have originated in this manner. The oölitic texture of the bauxite evidently led Doctor Hayes to believe that the ores have been precipitated from hot solutions, and consequently that the granitic type of ore represents replacement by hot solutions. As a result of a careful study of the oölitic ore and its relation to the granitic type of bauxite, it appears that this type of ore has been developed in place from the granitic or amorphous ore by the action of percolating water of the ordinary surface water type. The development of the pisolites is believed to be by a process similar to the formation of concretions. Conclusions are based on the following facts of observation: (1) Complete gradation may be observed between the granitic type of bauxite and the oölitic type, the oölites appearing in the granitic bauxite and becoming more numerous until the granitic texture is entirely obliterated. (2) The gradation from a textureless porous kaolin to oölitic bauxite can be observed,—the oölites beginning in very small individuals and becoming larger and more numerous as the ore is approached. (3) The relative position of the granitic and oölitic type of ore, the latter usually above the former where the surface solutions have been more active. (4) Microscopic examination of the oölitic ore reveals abundant feldspar forms in the oölites. (See Fig. 3.) These feldspar forms usually consist entirely of microcrystalline gibbsite, and are very striking indeed under the microscope, some of them even showing traces of the twinning of the plagioclase feldspars. (5) The solu-

tion and redeposition of the bauxite is possible chemically, as this material is soluble in acid solution. This texture was not observed to develop in non-bauxite kaolin which is not soluble. This texture is common to bauxite deposits the world over and is taken as one of the principal means of identification of bauxite. This makes it probable that the texture is due to some common and widely prevalent set of conditions and processes. (6) A pisolitic structure very similar to that of the bauxite has developed under similar conditions in certain iron ores of eastern Cuba, known to be the residual products of weathering in situ of serpentine rock. (7) The presence of heavy accessory minerals common to the syenite, evenly distributed throughout the pisolitic type of bauxite is perhaps the most convincing evidence that this type of ore has developed from the syenite in place.

EXTREME WEATHERING OF BASIC IGNEOUS ROCKS PRODUCING LATERITES AND LATERITIC IRON ORES

The processes and results of extreme weathering of basic igneous rocks are qualitatively the same as in acid igneous rocks; quantitatively they may be distinctly different, due to the higher content of iron in the product of the basic igneous rocks.

When basic rocks are altered under reducing conditions which permit the removal of iron in the ferrous form, the resulting product, whether a residual clay or a bauxite, is essentially like the products from acid igneous rocks, low in iron. On the other hand, when the alteration takes place under oxidizing conditions, the iron is retained with the residual clays and bauxites, giving a distinctly ferruginous residual clay or laterite, essentially different in iron content from the products of the less ferruginous acid igneous rocks. It is believed that the higher content of basic feldspars and of the easily altered ferromagnesian minerals, such as pyroxene and olivine, admits of more rapid alteration in these basic rocks, and, by developing a porous texture, due to the removal of all or part of these constituents, admits of more complete alteration of the residual clays to the hydrous-aluminum oxide, characteristic of laterites.

Extreme weathering of basic igneous rocks is well illustrated by the lateritic iron ores of northeastern Cuba, developed by the weathering of the serpentine rock, itself an alteration product

of a preëxisting peridotite. The following account of these deposits is based on detailed quantitative study.

Laterites of Northeastern Cuba [1]

Occurrence and nature of the deposits. The deposits constitute a nearly horizontal mantle from a few inches to nearly 80 feet thick over the serpentine country rock on a plateau 1,500 to 2,000 feet above sea level. Lack of active erosion has permitted the accumulation in situ of this thick mantle of decomposition products. The serpentine rock grades rather sharply into its alteration product, which, in the immediate neighborhood of the serpentine, is a greenish yellow material of a soft earthy texture and high porosity, consisting of a mixture of clay and hydrous iron oxide, with minute remnants of the serpentine which give the greenish tinge. This grades up into iron ore with a yellow color and of a fine-grained earthy texture. The porosity of the ore is remarkably high,—70 to 80 per cent of its volume, as shown by many tests. This ore may contain as high as 50 per cent moisture in place in the ground. The ore has the property of standing in vertical walls in the sides of open cuts and preserves the smallest pick marks and scratches for many months. Within 12 to 15 feet of the surface the ore begins to take on a darker and more reddish color and minute granules or shot, consisting of hematite and magnetite, appear. The granules have radial and concentric textures. The red color and the amount and size of shot increase on approaching the surface and in some deposits the upper foot or two feet of ore consists of a dark colored cemented mass of shot locally known as *plancha*. A face of ore in the side of a test pit or open pit presents a remarkably uniform gradation in color from a dark reddish ore at the top to the light yellow ore at the bottom. The change from yellow ore to reddish ore and the change from smooth earthy texture to somewhat coarser texture, with abundant shot, is accompanied by a change from limonite to hematite and magnetite. Chemical analyses also show that clay

[1] Leith, C. K., and Mead, W. J., Origin of the iron ores of central and northeastern Cuba: Trans. Am. Inst. Min. Engrs., vol. 42, 1912, pp. 90–102; Additional data on origin of lateritic iron ores of eastern Cuba: Bull. 103, Am. Inst. Min. Engrs., July, 1915, pp. 1377–1380.

Kemp, J. F., The Mayari iron ore deposits, Cuba: Bull. Am. Inst. Min. Engrs., Feb., 1915, pp. 129–154.

present in the ore immediately above the serpentine rock is altered to bauxite or gibbsite on approaching the surface.

The chemical changes in alteration of serpentine to iron ores. In the following table is a series of partial chemical analyses of samples taken every foot from the surface to the serpentine rock at a depth of 29 feet.

TABLE V. CHEMICAL ANALYSES SHOWING ALTERATION OF SERPENTINE ROCK TO IRON ORE IN THE MAYARI DISTRICT, CUBA

Analyses supplied by Spanish-American Iron Company

Depth, feet	SiO_2	Al_2O_3	Fe_2O_3	Fe	MgO	Cr	Ni+Co	P	S	H_2O+	Total
0– 1	2.58	15.71	66.20	46.37		.92	.38	.016	.12	10.20	96.126
1– 2	2.38	20.81	64.70	45.34		.96	.33	.022	.12	10.63	99.952
2– 3	1.60	17.43	68.40	47.81		.96	.42	.018	.14	9.15	98.118
3– 4	1.42	14.23	68.70	48.09		1.04	.36	.019	.16	9.50	95.429
4– 5	1.56	8.47	70.60	49.46		1.27	.61	.016	.17	10.14	92.836
5– 6	2.90	10.24	72.35	50.56		1.66	.84	.016	.20	10.96	99.166
6– 7	2.20	8.29	72.90	51.00		2.19	1.09	.007	.19	11.35	98.217
7– 8	2.68	4.92	71.85	50.28		2.19	1.15	.006	.14	11.57	94.506
8– 9	3.30	7.25	71.55	50.15		2.39	1.14	.006	.16	12.12	97.916
9–10	2.44	6.91	72.40	50.63		2.08	1.21	.005	.16	12.35	97.555
10–11	2.42	6.31	71.40	49.94		2.00	1.36	.005	.14	12.40	96.035
11–12	2.72	7.05	70.55	49.46		2.08	1.31	.004	.15	12.40	96.264
12–13	2.56	6.77	70.20	49.08		1.62	1.37	.004	.10	13.50	96.124
13–14	2.52	6.23	70.55	49.46		1.85	1.41	.005	.14	13.12	95.825
14–15	2.76	6.58	71.85	50.22		1.89	1.38	.007	.21	12.45	97.127
15–16	2.78	6.53	70.00	48.98		2.16	1.33	.007	.19	12.35	95.347
16–17	2.98	6.43	69.80	48.84		2.19	1.42	.007	.19	12.57	95.587
17–18	3.20	5.53	70.45	49.32		2.00	1.35	.007	.15	12.90	95.587
18–19	3.66	6.51	69.20	48.42		2.43	1.34	.005	.06	12.73	95.935
19–20	6.84	8.49	63.35	44.32		2.51	1.36	.004	.08	12.80	95.434
20–21	7.44	5.13	66.55	46.58		2.27	1.57	.003	.09	12.45	95.503
21–22	8.46	4.99	57.80	40.47	.00	2.16	1.47	.006	.08	12.71	87.676
22–23	11.04	8.38	62.10	43.49	.00	1.85	1.74	.002	.09	14.07	99.272
23–24	15.86	4.70	63.90	44.62	.00	2.19	1.57	.003	.09	11.73	100.043
24–25	17.40	4.00	62.90	40.00	.50	1.85	1.43	.003	.12	11.64	99.843
25–26	22.54	4.57	50.25	35.12	6.49	1.89	1.80	.002	.06	13.65	101.252
26–27	28.60	4.18	32.85	23.00	18.23	1.12	1.43	.003	.09	13.45	99.953
27–28	35.64	2.33	18.25	12.78	27.35	0.77	1.35	.001	.06	14.23	99.981
28–29	39.80	1.39	10.14	7.10	33.69	0.20	0.97	.001	.06	13.31	99.561

In Fig. 6 the change in chemical composition from the serpentine rock upward through the ore to the surface is expressed by means of a composite "straight-line" diagram. Each analysis in turn is compared with the analysis of the serpentine rock at the

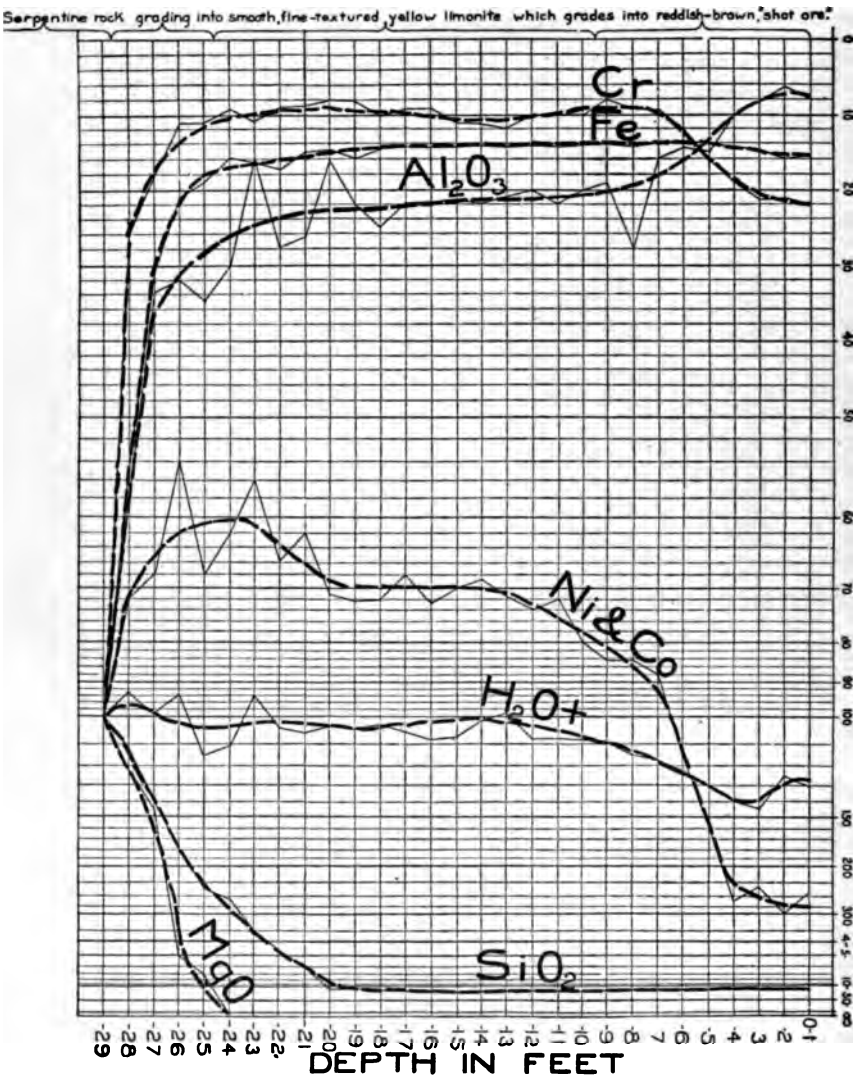

Serpentine rock grading into smooth, fine-textured yellow limonite which grades into reddish-brown, "shot ore."
Cr
Fe
Al_2O_3
Ni & Co
H_2O+
MgO
SiO_2
-29 -28 -27 -26 -25 -24 -23 -22 -21 -20 -19 -18 -17 -16 -15 -14 -13 -12 -11 -10 -9 -8 -7 -6 -5 -4 -3 -2 -1 0
DEPTH IN FEET

Fig. 6. Representing the chemical change in the katamorphism of serpentine rock to iron ore, eastern Cuba.

base, and the several points on the lines of the diagram are connected to form curves, which indicate the relative losses and gains of constituents. This diagram shows the early and complete loss of magnesia, large loss of silica, marked increase of alumina, iron and chromium, and relative loss of cobalt and nickel. The diagram further shows loss of iron relative to alumina near the sur-

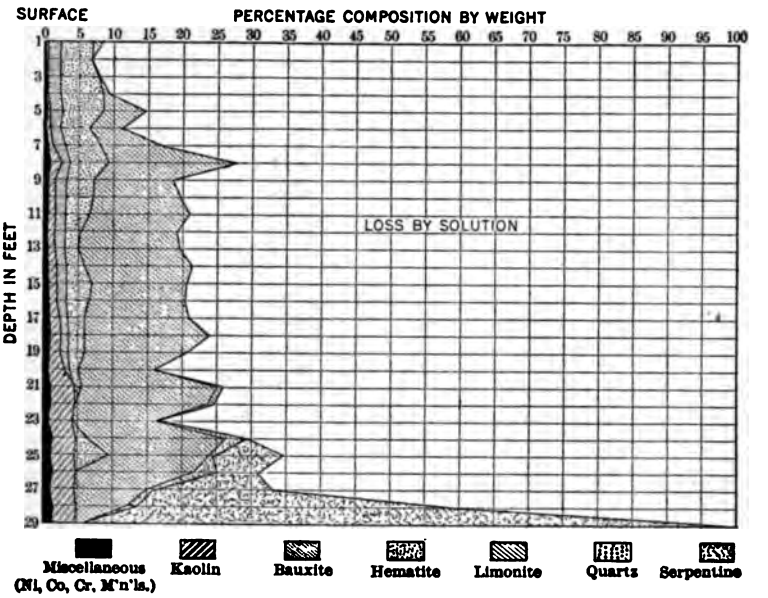

SURFACE
PERCENTAGE COMPOSITION BY WEIGHT
0 5 10 15 20 25 30 35 40 45 50 55 60 65 70 75 80 85 90 95 100
1 3 5 7 9 11 13 15 17 19 21 23 25 27 29
DEPTH IN FEET
LOSS BY SOLUTION
Miscellaneous (Ni, Co, Cr, M'n'ls.)
Kaolin
Bauxite
Hematite
Limonite
Quartz
Serpentine

FIG. 7. Representing in terms of weight the mineralogical changes in the katamorphism of serpentine rock to iron ore, on the assumption that alumina has remained constant, eastern Cuba.

face and an increase of iron relative to alumina below. This indicates solution of iron at the surface and its deposition at depth, resulting in a downward concentration of iron. This is believed to be due to the reduction of iron through the action of plant remains and plant roots near the surface, its transportation downward in the ferrous condition, and its precipitation by oxidation on meeting laterally moving oxidizing waters below the zone of plant roots. This reduction and reprecipitation of iron through the agency of the organic material and plant roots is believed to be one of the causes for the change in the color of the ore. It may

also explain the development of the small granules or pisolites which are found only to the depth of the plant roots.

The leaching of cobalt and nickel near the surface suggests the possibility of the existence of cobalt-nickel veins in the serpentine, as in New Caledonia. Very little exploration has been carried down into the serpentine.

Mineralogical changes in the development of the ore. In Fig. 7 the gradation in mineralogical composition from the serpentine rock to the surface is shown in terms of weight, and on the assumption that alumina has remained constant during the alteration. The mineral composition of each sample was calculated from the chemical analysis in the foregoing table. This diagram illustrates the very rapid destruction of serpentine by the leaching of silica and magnesia, a change from kaolin to the hydrous oxides of alumina, and the development of hematite and magnetite from limonite on approaching the surface. While the mineral quartz has not been actually observed in the ores, calculation of the chemical analyses indicates a certain amount of this mineral in the zone immediately above the serpentine rock. It seems probable that some of the large amount of silica freed by the removal of magnesia is temporarily deposited as quartz or opal, which is later removed in solution as the alteration proceeds.

COMPARISON OF LATERITIC IRON ORES OF CUBA AND THE ARKANSAS BAUXITE DEPOSITS

These two types of alteration products, one developed from very basic igneous rocks, and one from acid igneous rocks, have interesting points of similarity. The Cuban ores have been developed from a serpentine rock, comparatively rich in iron, with a very low content of alumina. The bauxite has been developed from a nepheline syenite, having a high content of alumina and a comparatively low iron content. The first step, both in the alteration of the serpentine rock and the nepheline syenite, is the kaolinization of the aluminous minerals, loss of silica, magnesia, and alkalies, and oxidation of iron. This produces from the serpentine rock a porous limonitic mass with a small amount of kaolin, and from the nepheline syenite a porous mass of kaolin containing a small amount of hydrated iron oxide. Further alteration in both cases consists of a leaching of silica from the

kaolin, changing that mineral to a hydrated oxide of alumina, but yielding in one case a limonite, containing a small amount of bauxite, and in the other, a bauxite, containing a minor amount of limonite. Near the surface in both cases small pisolites begin to develop, consisting of iron oxide in the iron ores and of bauxite in the bauxite ores. In both the iron ores and the bauxite this pisolitic texture becomes more marked toward the surface, and at the surface develops a hard mass made up of relatively large concretions thoroughly cemented together. In the Cuban iron ores this constitutes what is known as the *plancha* or sheet deposit, and in the bauxite this surface phase is known as the *hard bauxite.*

OCCURRENCE AND FORMATION OF LATERITES IN GENERAL

The two most stable constituents under conditions of surface weathering are alumina and ferric iron. Under extreme conditions of weathering the residual products of rock decay consist essentially of ferric hydroxide and aluminum hydroxide, in other words, limonite and bauxite. The relative amounts of limonite and bauxite in residual products of rock decay obviously depend on the relative amount of iron and alumina in the parent rock. It is true that alteration under reducing conditions may permit the removal of the iron in solution, leaving a product of alteration relatively richer in alumina, but the more general case is one in which the iron is oxidized and remains in the form of limonite or other ferric hydrate, with the residual material. The Arkansas bauxites represent one extreme in which a rock rich in alumina and containing very little iron yields a comparatively pure bauxite, while the lateritic iron ores of Cuba represent the other extreme in which a rock fairly rich in iron, and containing very little alumina, yields a limonitic iron ore containing a small amount of bauxite.

The products of weathering lying between these two extremes, that is, containing various proportions of ferric hydroxide and alumina hydroxide, with varying amounts of free silica and clay, have been called "*laterites.*" [1] Laterites occur rather widely through the tropical and semi-tropical regions, both as residual products of rock decay and as transported sediments. Laterites have been described as resulting from alterations in situ of various

[1] For an excellent summary of laterites and laterization, see Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, pp. 469–477.

types of both basic and acid igneous rocks, of lavas and volcanic ash, and of limestone and dolomite. There appears to be no reason why argillaceous sediments, such as the shales, should not yield to laterization, unless it is that their fine texture prevents easy flow of water. In general, the residual laterites have less of impurities than the transported laterites, which may contain considerable admixtures of free silica and clay.

There seems to be some question as to whether the aluminum hydroxides are developed directly from the silicates of the altered rock or whether the process involves a preliminary kaolinization of the minerals, followed by the desilication of the kaolin. In the two cases described in detail in this text there is evidence of the intermediate hydrous aluminum silicate phase, preliminary to the formation of the aluminum hydroxide. Since the development of hydrous aluminum silicates from the aluminous minerals of igneous rocks is such a common method of alteration, it seems reasonable to look upon the development of aluminum hydroxide in rock decay as simply one step further in the alteration. It is of course possible that this alteration may go on so rapidly as to obscure the intermediate hydrous silicate stage.

CHAPTER IV

HYDROTHERMAL KATAMORPHISM OF IGNEOUS ROCKS

MINERAL AND CHEMICAL CHANGES IN HYDROTHERMAL KATAMORPHISM OF IGNEOUS ROCKS

Rock alterations of a distinctive type which have been accredited to hot solutions occur locally along fissures and other openings in igneous rocks and near the surface of flows. These have been studied particularly in connection with certain metallic ores which have been formed by hot solutions. Such alterations near the surface and to moderate depths show some of the features of katamorphism and are described in this section, but at higher temperatures and pressures, ordinarily correlated with greater depth, such changes are more anamorphic in their nature and may be more conveniently described in connection with anamorphism. The anamorphic and katamorphic alterations by hot solutions more or less overlap.

The process of alteration by hot solutions at the surface can be well observed around live and recently extinct volcanoes. Here are areas of reddish, brown, and yellowish rocks, soft and disintegrated, consisting of quartz, iron oxide, calcite, barite, fluorite, kaolin, nontronite, jarosite, alunite, gypsum, and other sulphates, exceptionally adularia and zeolites, all of which are obviously formed by the alteration of the lavas under the influence of the carbonic, sulphuric, and other acids contained in the volcanic solutions, both aqueous and gaseous. The quartz is usually fine-grained or chalcedonic and where iron stained it is jaspery. The minerals may be locally segregated in patches. Some of the minerals are like those developed by weathering, others, like the sulphate minerals and adularia, are more or less distinctive of alterations by hot solutions. Chemically the change is not

unlike weathering, particularly in the loss of soda and the gain of water, but on the whole the change shows much more irregularity of behavior of silica and the bases than weathering. Sulphur is often gained.

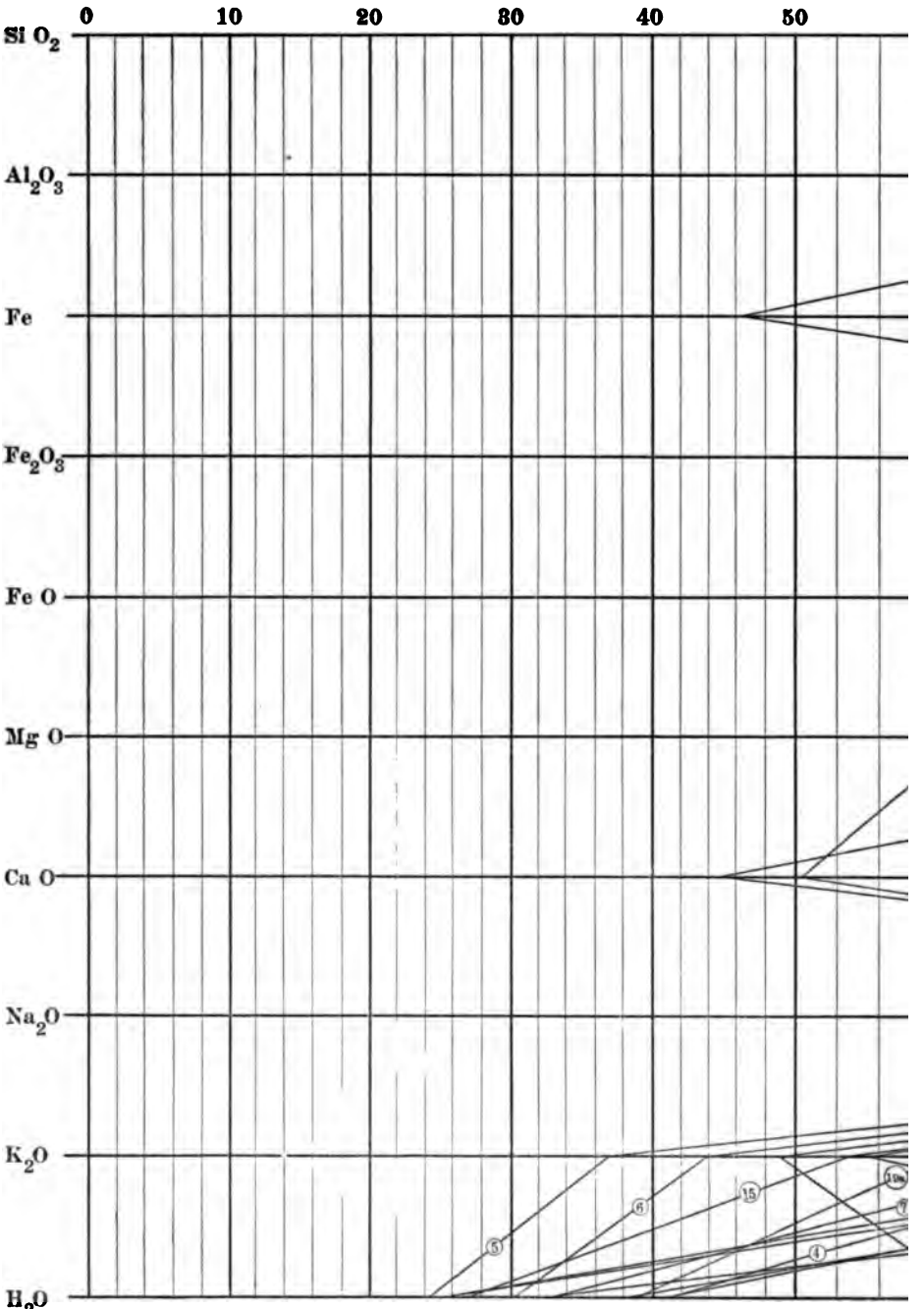
More or less overlapping this surface alteration, but more characteristic of moderate depths beneath the surface, are two other types of alteration, one characterized by the development of quartz, sericite, and adularia, and a type characterized by the development of chlorite and epidote. Both types contain pyrite and other sulphides, and more or less carbonates. Albite and other minerals are locally developed. The chloritic type has been called *propylite*. Both of these phases are well cemented and crystallized. Both are associated with important ores. Frequently the quartz-sericite alteration predominates near the vein and the chloritic alteration further away from the vein, suggesting that the rocks have acted as a screen or filter, changing the nature of the solutions penetrating into the rock from the vein. Adularia often occurs at shallower depths than sericite, according to Lindgren.[1] Carbonates and albite on the whole are more characteristic of intermediate than shallow depths.

The most distinctive mineral feature in common in these alterations is the extensive development of potash minerals, sericite, adularia, and alunite. The fact that clay is often associated with alterations of this kind is presumably often due to later weathering by acid meteoric solutions from the surface, though it has been shown in some cases that hot waters may develop clay. It is also to be noted that sericite is not formed exclusively by hot solutions, because this mineral also develops in ordinary weathering. Possibly also some alunite is formed this way, though it is usually ascribed to reactions of hot solutions from below with acid solutions from above.

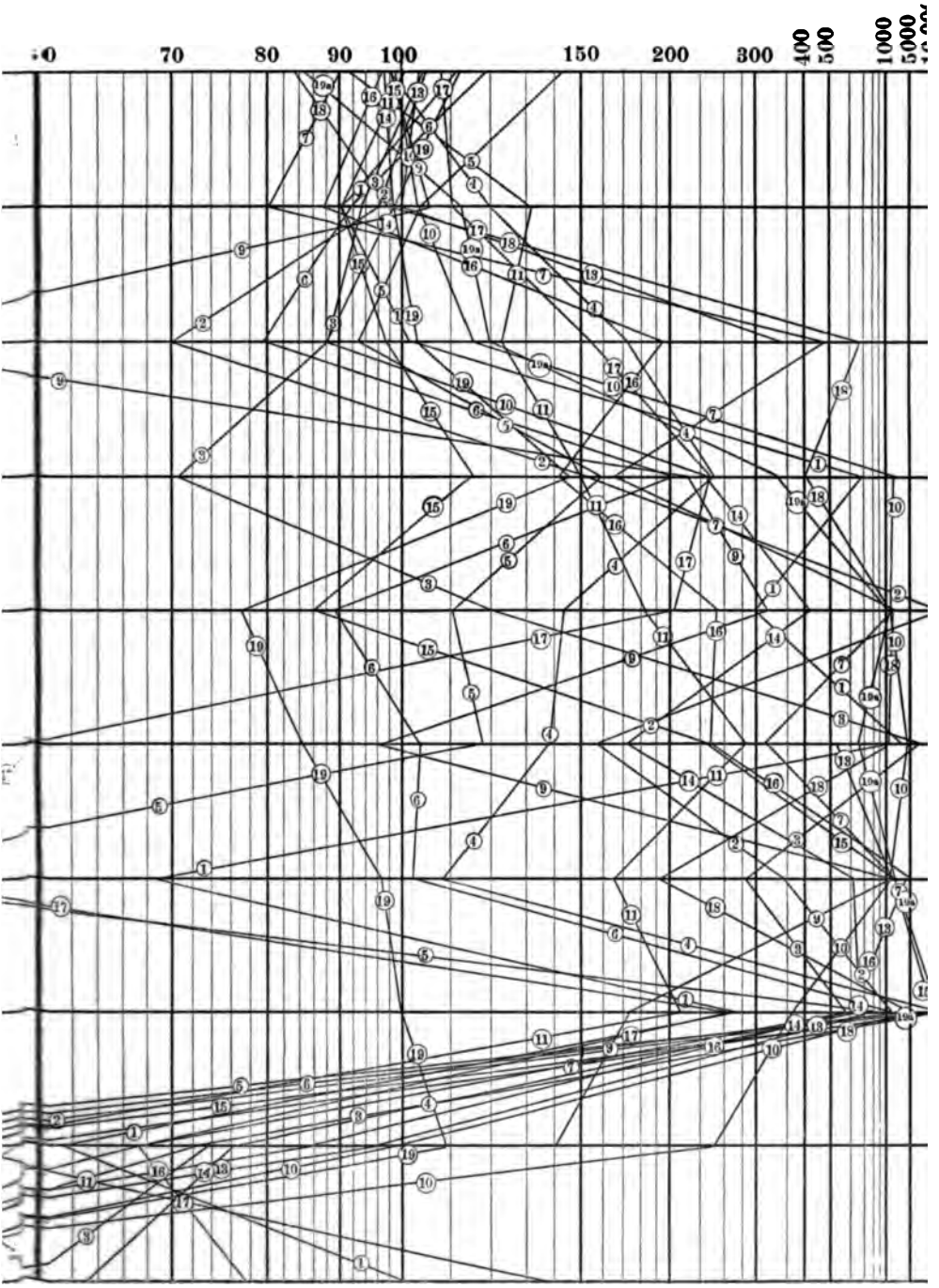
Chemically the changes roughly follow those of ordinary katamorphism of acid igneous rocks but with certain distinctive differences. In weathering the bases behave rather uniformly, while in the thermal alteration there is much irregularity in their gains and losses. Magnesia, soda, and ferrous and ferric oxide are usually lost, and potassa, sulphur, carbon dixoide, and water

[1] Lindgren, Waldemar, Mineral deposits, p. 457, New York, 1913.

PLATE V

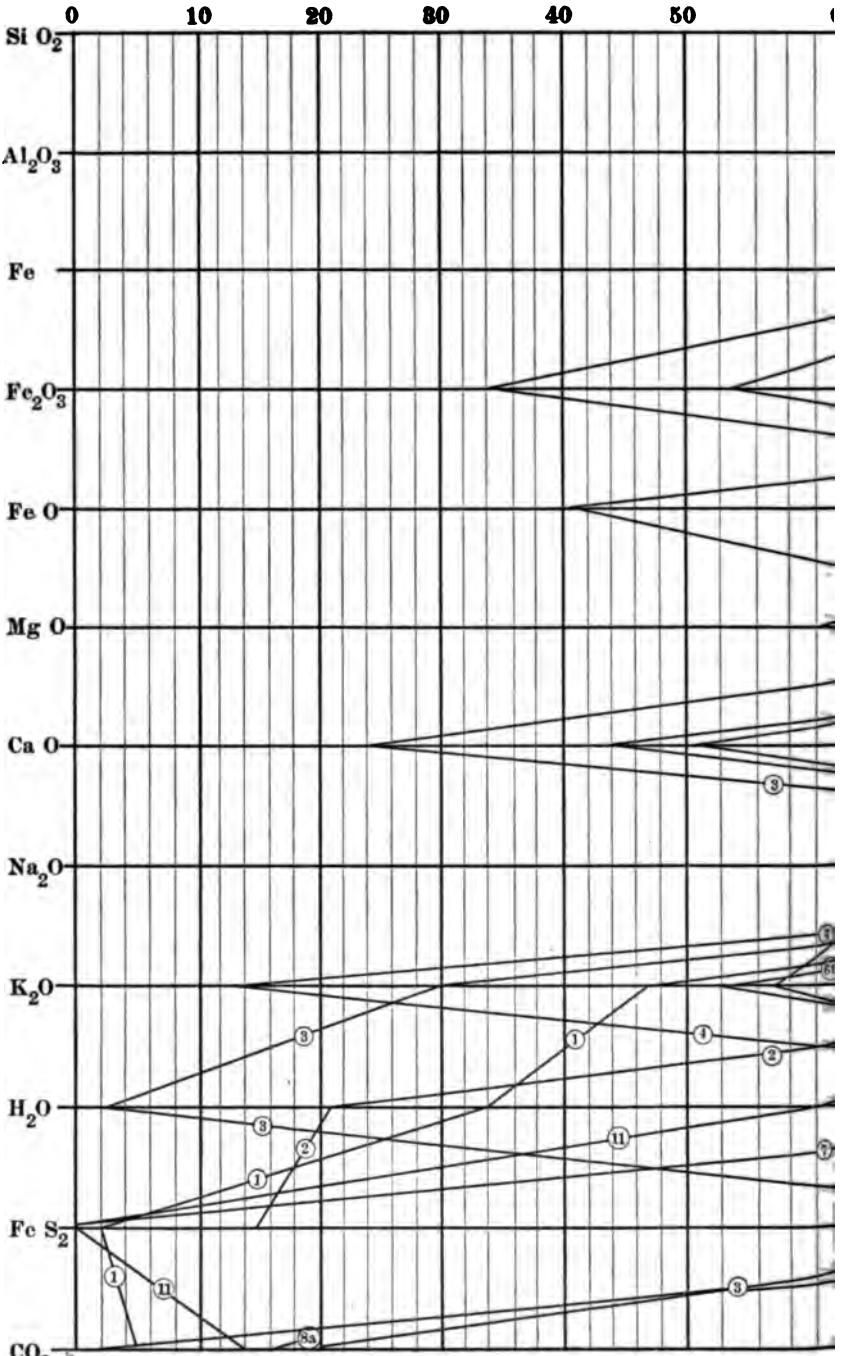


Representing the gains and losses of constituents in the katamorphis
Each broken line represents comparison

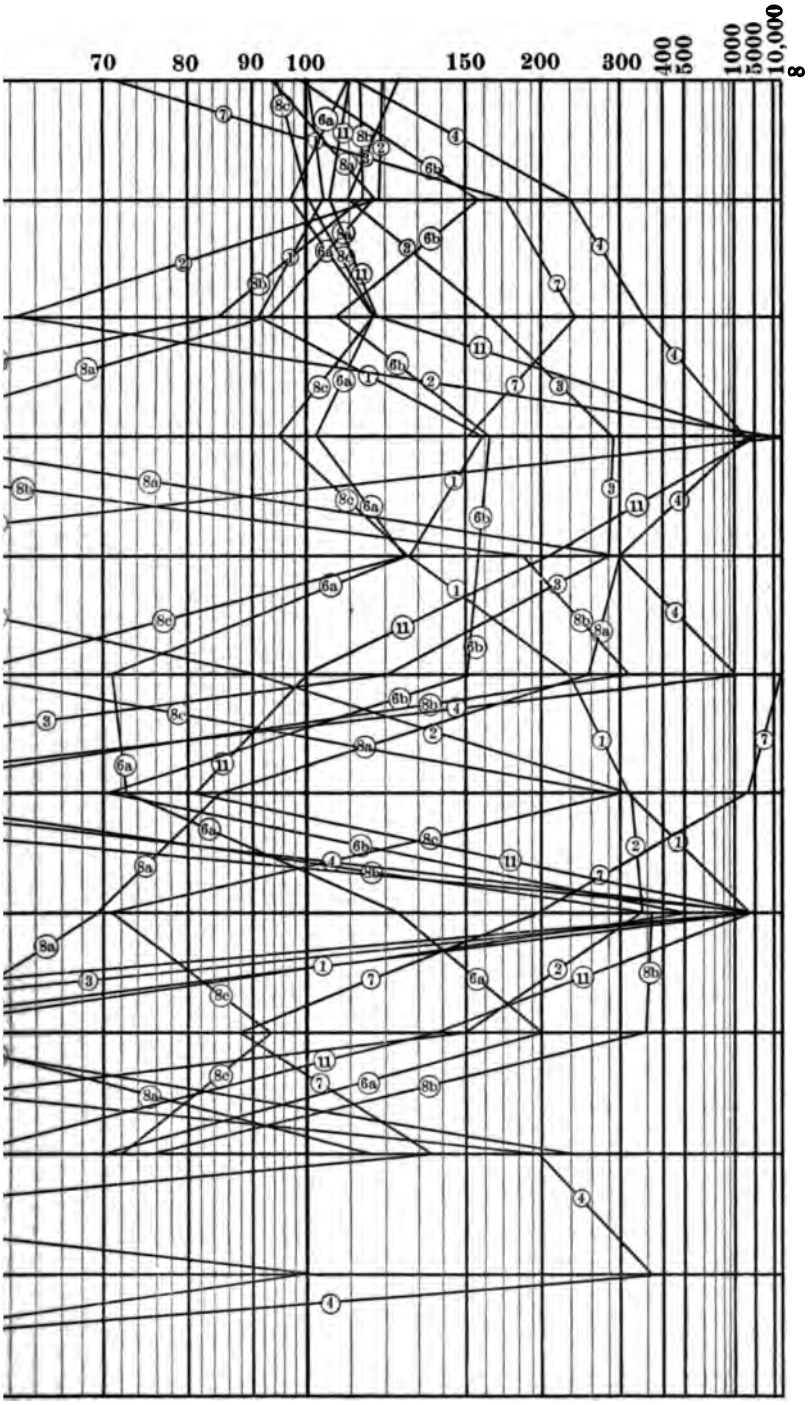


cid igneous rocks by thermal solutions at shallow and intermediate depths.
air of analyses, numbered for identification.

PLATE VI



Representing the gains and losses of constituents in the katamorphis
mediate depths. Each broken line represents compos



of basic igneous rocks by thermal solutions at shallow and inter-
m of a pair of analyses, numbered for identification.

are gained. The ferrous iron enters into the sulphide minerals. Silica may be gained or lost with reference to alumina, whereas in weathering it is uniformly lost. Sulphur is gained. Carbon dioxide is introduced in the cases which involve the development of carbonates.

Hydrothermal alterations of acid and basic igneous rocks are summarized on Plates V and VI. Following is a list of references to analyses platted on Plate V.

1. Granitic rock, Cripple Creek, Colorado. Lindgren, W., and Ransome, F. L., Prof. Paper 54, U. S. Geol. Survey, p. 194.
2. Latite phonolite, Cripple Creek district, Colorado. Lindgren, W., and Ransome, F. L., Prof. Paper 54, U. S. Geol. Survey, p. 189.
3. Monzonite porphyry, Bingham district. Boutwell, J. M., Prof. Paper 38, U. S. Geol. Survey, p. 178.
4. Granitic rock, Willow Creek district, Idaho. 18th Ann. Rept., U. S. Geol. Survey, pt. 3, p. 640.
5. Granodiorite, Placer County, California. Trans. Am. Inst. Min. Engrs., vol. 30, 1901, p. 666.
6. Granodiorite, Nevada County, California. Trans. Am. Inst. Min. Engrs., vol. 30, 1901, p. 666.
7. Granite, Pilbara gold field, Australia. Lindgren, W., Econ. Geol., vol. 1, 1906, p. 540.
9. Rhyolite porphyry, Butte, Montana. Weed, W. H., Prof. Paper 74, U. S. Geol. Survey, p. 90.
10. Dacite, Goldfield, Nevada. Ransome, F. L., Prof. Paper 66, U. S. Geol. Survey, p. 181.
11. Hornblende dacite, Hauraki gold field, New Zealand. Finlayson, A. M., Econ. Geol., vol. 4, 1909, p. 638.
13. Quartz monzonite, Butte, Montana. Weed, W. H., Prof. Paper 74, U. S. Geol. Survey, p. 88.
14. Granite, Pittsmont Mine, Butte, Montana. Unpublished.
15. Quartz monzonite, San Francisco district, Utah. Butler, B. S., Geology and ore deposits of the San Francisco region, Utah: Prof. Paper 80, U. S. Geol. Survey, p. 76.
16. Quartz monzonite, San Francisco region, Utah. Butler, B. S., Geology and ore deposits of San Francisco region, Utah: Prof. Paper 80, U. S. Geol. Survey, p. 76.
17. Porphyry, Rio Tinto, Spain. Lindgren, W., Mineral Deposits, p. 604.
18. Monzonite, Tintic, Utah. Lindgren, W., Mineral Deposits, p. 525.

19. } Hornblende-mica andesite (earlier andesite), Tonopah, Nevada.
19a } Lindgren, W., Mineral Deposits, p. 452.

Following is a list of references to analyses platted on Plate VI.

1. Diorite, Willow Creek district, Idaho. 20th Ann. Rept., U. S. Geol. Survey, pt. 3, p. 219.
2. Diorite porphyry, Breckenridge district, Colorado. Ransome, F. L., Prof. Paper 75, U. S. Geol. Survey, p. 96.
3. Amphibole schist, Placer County, California. Trans. Am. Inst. Min. Engrs., vol. 30, 1901, p. 666.
4. Diabase, Grass Valley, California. Trans. Am. Inst. Min. Engrs., 1901, p. 666.

6a } Hornblende andesite, Hauraki gold field, New Zealand. Finlayson,
6b } A. M., Econ. Geol., vol. 4, 1909, p. 637.

7. Diabase, Comstock Lode. Becker, G. F., Mon. 3, U. S. Geol. Survey, p. 153.

8a. Trap rock, Keweenaw Point, Michigan. Mon. 52, U. S. Geol. Survey, p. 583.

8b. Trap (melaphyre), Keweenaw Point. Pumpelly, R., Proc. Am. Acad. Arts and Sci., vol. 13, 1878, p. 293.

8c. Trap, Keweenaw Point. Pumpelly, R., Proc. Am. Acad. Arts and Sci., vol. 13, 1878, p. 285.

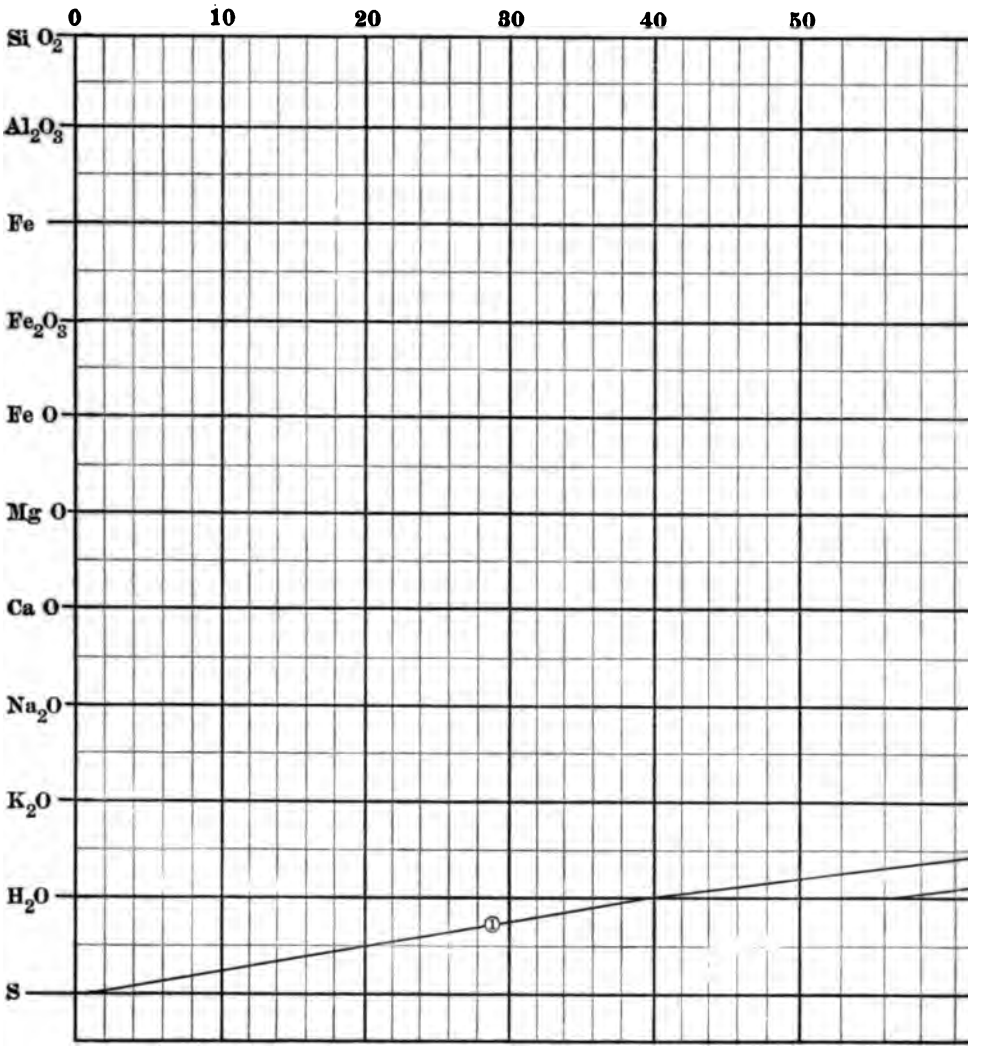
11. Andesite (later), Tonopah, Nevada. Lindgren, W., Mineral Deposits, p. 454.

A comparison of these plates with plates showing ordinary weathering of acid and basic igneous rocks (Pls. II and IV), will indicate better than words the essential chemical differences in the alterations.

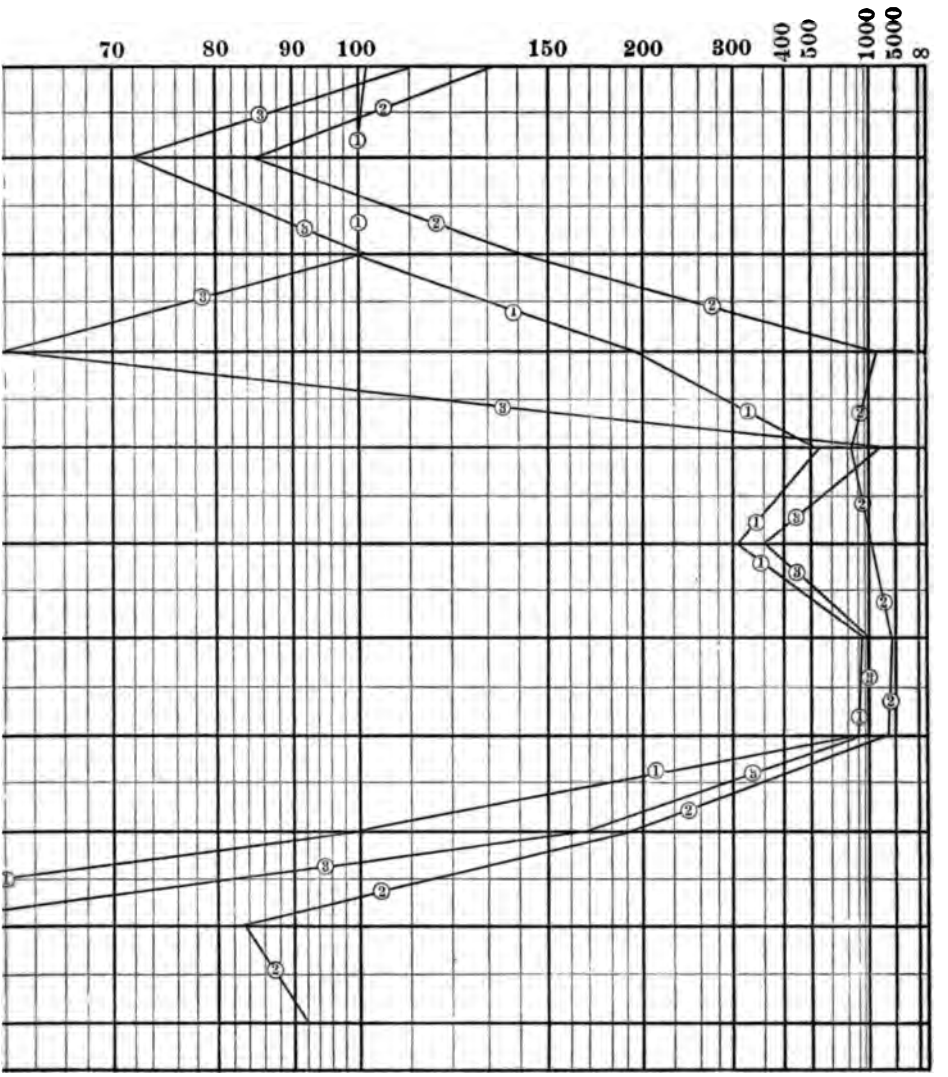
A specific case of sericitic alteration is illustrated in Plate VII representing the hot water alterations in the granite adjacent to the copper veins at Butte. Adjacent to the veins the granite is changed to a sericitic, silicified and pyritized granite. Pyritization conspicuously attacks the ferromagnesian minerals. Near the surface weathering has affected the sericitic phases, with the result that clay takes the place of sericite and residual feldspars, and iron oxide develops from the pyrite. The total iron, alumina, silica, and potash remain substantially without change during the sericitic alteration.

The hydrothermal alterations of igneous rocks, above described, are essentially katamorphic in their destruction of primary silicates, and the development of secondary hydrous silicates and carbonates. At the surface the process softens and disintegrates the rock mass. Below the surface the newly formed minerals thoroughly cement the rock, and in this respect only is the change

PLATE VII



Representing gains and losses of constituents in wall rock alterations in
wall rock. No. 2. Later alteration of sericitized wall rock by cold
ing of acid igneous rocks for comparison with No. 1 and No. 2.



Pittsmont Mine, Butte, Montana. No. 1. Hydrothermal alteration of
ending acid solutions. No. 3. Composite of average changes in weather-

more nearly analogous to the anamorphic process of cementation than katamorphism. The same solutions which develop these changes in the igneous rocks cement and anamorphose adjacent sedimentary rocks. This is but one of many illustrations, which could be cited, of the simultaneous progress of both katamorphic and anamorphic changes under the same environment.

At great depths and high temperatures igneous rocks undergo alterations by hot solutions which are not katamorphic in their general nature. Neither are they distinctively anamorphic, but, as they occur under distinctively anamorphic conditions and in anamorphic associations, they are described in Chapter IV of Part II.

VOLUME CHANGES IN HYDROTHERMAL KATAMORPHISM OF IGNEOUS ROCKS

The hydrothermal alterations at and near the surface may result in increased porosity and actual disintegration, and mechanical expansion resulting in a volume increase. At greater depth the rock pressure appears to be sufficient to prevent expansion and the alterations may proceed at constant volume. This is believed to be the case in the alteration of the quartz monzonite at Butte, Montana. It is conceivable that at great depths hydrothermal alteration with formation of heavy silicates may involve decrease in volume.

CHAPTER V

KATAMORPHISM OF SEDIMENTS

The ordinary residual products of weathering of igneous rocks, that is, clays and soils, are themselves in the nature of end-products of katamorphism described on preceding pages and ordinarily undergo no further alteration, though under certain conditions they may be laterized. It would be expected that these materials when transported and deposited as clays and sands would suffer little further change of a katamorphic nature when cemented to shales and sandstones and again exposed to weathering.

The clear quartz sands suffer slight solution, as shown by etched surfaces, but in proportion as they contain original silicates, they may further decompose in the same manner as the igneous rocks.

The clays and shales usually contain some proportion of silicate mineral not yet broken down and also certain secondary silicates formed during cementation, and as long as these are present there is of course a tendency for their destruction by katamorphism in the manner described above for the igneous rocks. The iron in the muds, during decomposition and the subsequent cementation, is usually reduced to ferrous form as indicated by the green and grayish colors of the sediment. It is usually in minerals of a chloritic nature. When exposed again to weathering there is a tendency for the oxidation of the iron, accompanying the breaking down of silicates, giving reddish, brownish, and grayish weathered surfaces. Where a shale is made up largely of particles derived from the mechanical disintegration of igneous rocks or from volcanic dust, katamorphism may be in kind and scale like that affecting the igneous rocks. Where the shale is made up of materials which have been extensively decomposed, the possibilities for further alteration are small. Even where the shales contain abundant silicates they are as a whole slower in decomposing than

igneous rocks, because of their extremely fine grain which does not permit of the easy circulation of water. Where further indurated into slates and schists alteration is still slower. Hot waters locally play a part in katamorphism of shales (See p. 114.)

The limestones constitute the third great section of the sediments. Resulting, as they do, from processes of katamorphism, and being in the nature of end-products of such alteration, it might be supposed that they are adapted to environment and would not readily change when subjected to further katamorphism, but in this respect they differ widely from the mechanical sediments, sandstones, and shales, or from residual soils, in that they are readily affected by weathering. The lime and magnesia carbonates are not stable under weathering conditions. In that sense they do not seem to be permanently adapted to the environment under which they are formed. As long as they remain under water they are protected, though even here limestone may be dolomitized. (See pp. 149–151.) When exposed again at the surface they go rapidly into solution and are again carried away to be deposited under water. Their habit is distinctly subaqueous and they are destroyed when long exposed near or at the surface. The result of the weathering of a carbonate formation is usually a mass of clay containing chert fragments, and more or less hydrous iron oxide. Some of the clay may be reduced to bauxite. Where the iron is in considerable amount in the limestone or dolomite in the form of carbonate, or oxide, or even sulphide, the residual product after the elimination of the calcium and magnesium carbonates may be rich enough in iron to be called an iron ore. The brown ores or limonites forming so abundantly over the limestones and dolomites of the southern Appalachians consist of irregular residual masses of limonite in a matrix of clay, formed by ordinary processes of weathering. Of course such an ore requires washing to eliminate clay, to make it usable. During the alteration there is a tendency for the iron and clay to separate in concentric forms, giving granules, and shells of iron oxide filled with clay. Iron oxide seems to work outward during this alteration.

In Plate VIII, 4 pairs of analyses showing weathering of limestones and dolomites are represented graphically to show the relative losses and gains of constituents. Following is a list of the references to analyses platted:

1. Limestone, Wythe County, Virginia. Watson, T. L., Bull. 1, Virginia Geol. Survey, p. 98.
2. Galena limestone, southwestern Wisconsin. Grant, U. S., Bull. 14, Wis. Geol. Survey, pp. 15–16.
3. Knox dolomite, Morrisville, Alabama. Russell, I. C., Bull. 52, U. S. Geol. Survey, 1889, p. 25.
6. Limestone, Batesville district, Arkansas. Penrose, R. A. F., Jr., Ann. Rept., Ark. Geol. Survey, vol. 1, 1890, p. 180.

Cherty iron carbonates and ferrous silicates form extensive sedimentary formations in the Lake Superior iron region. Oxidation produces limonite, while very subordinate quantities of the

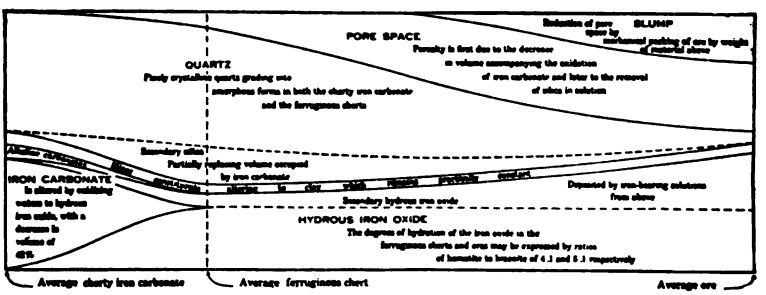
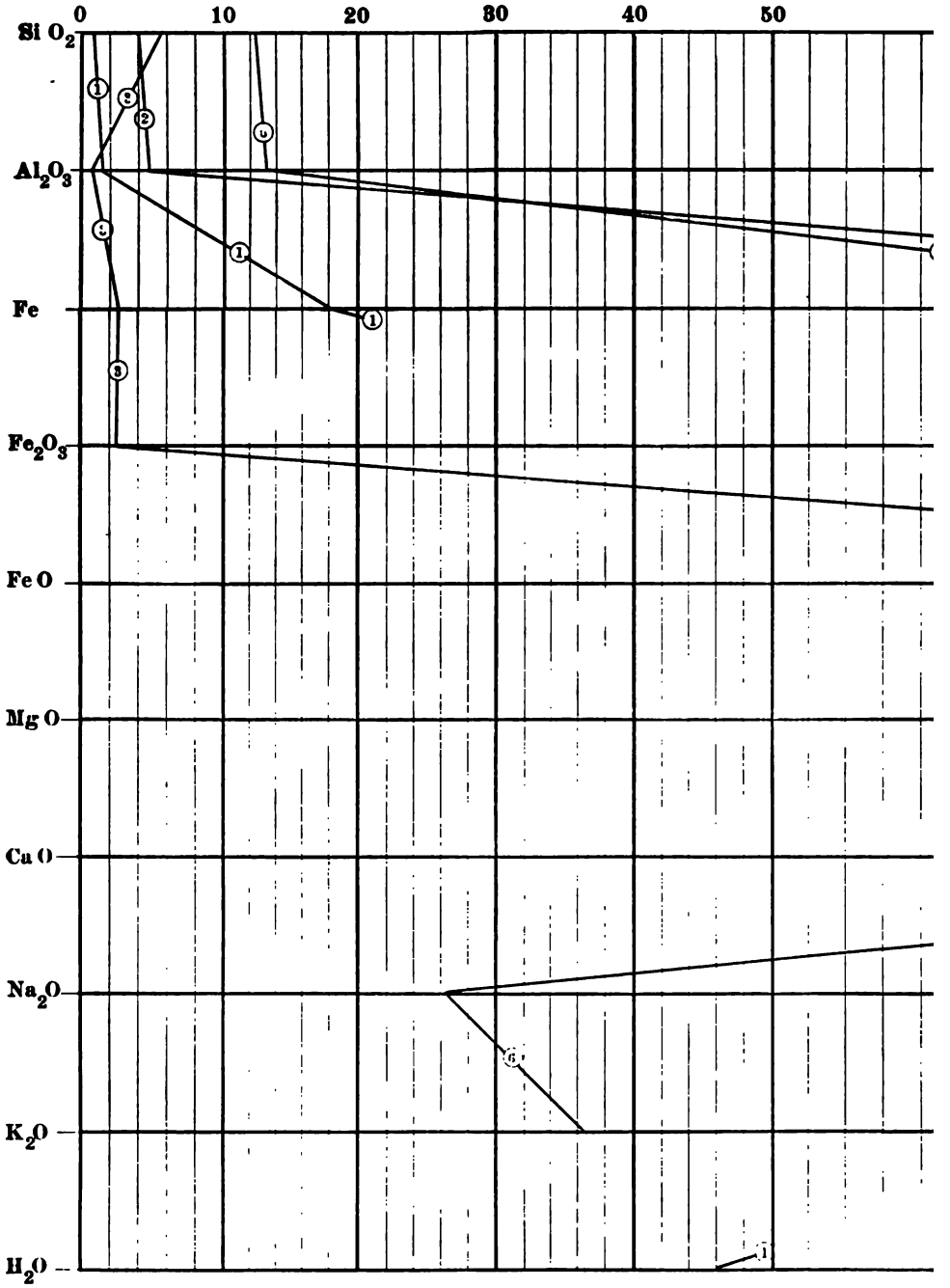


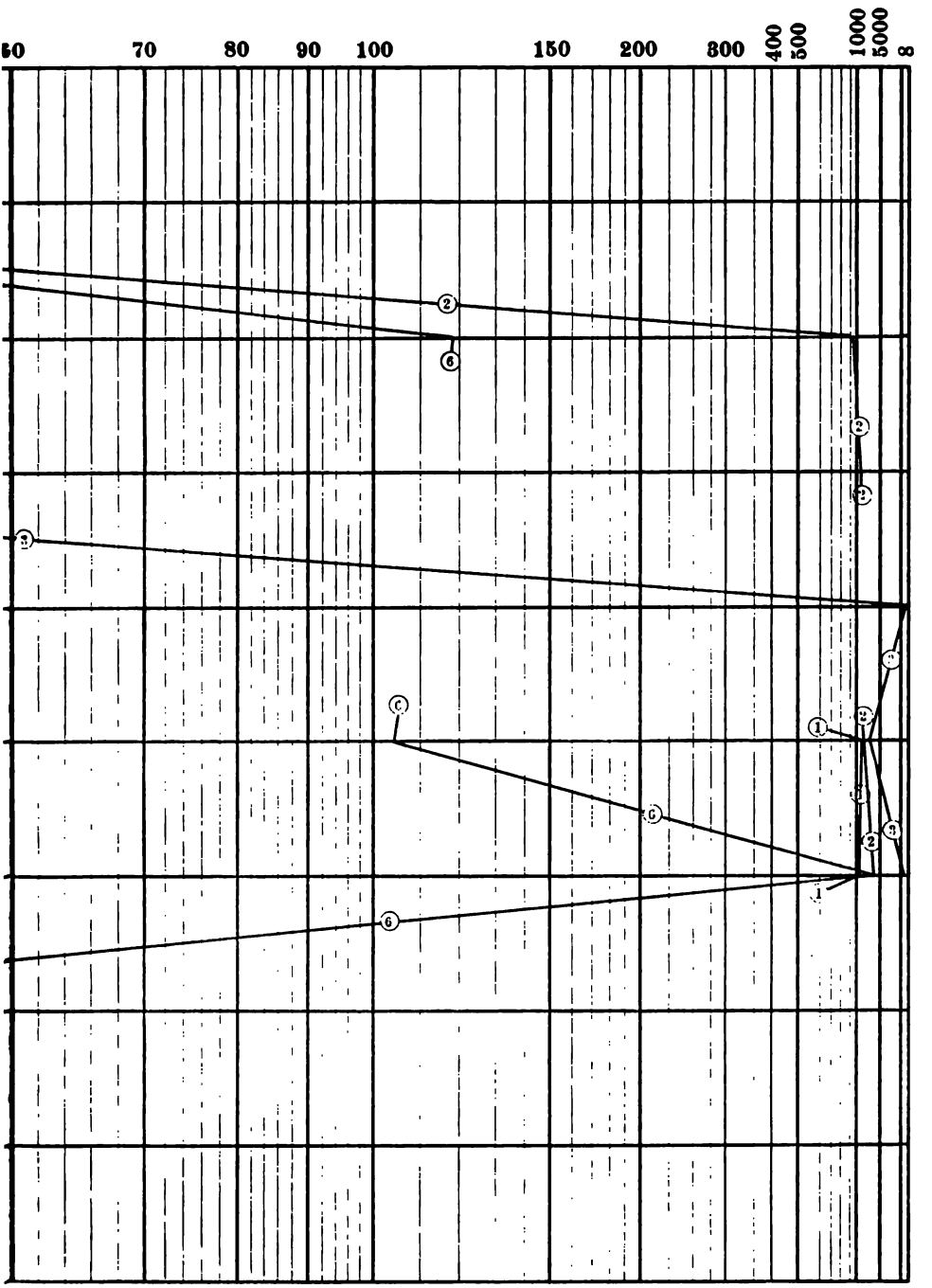
FIG. 8. Representing in terms of volume a general case of the weathering of iron carbonate from the Lake Superior region. After Van Hise and Leith (Mon. 52, U. S. Geol. Survey).

iron are carried away as bicarbonate or sulphate. The silica of the silicate is removed in solution. The cherty quartz, so commonly interbedded with the original carbonate or silicate, is slowly removed. Oxidation is complete long before any appreciable amount of quartz is removed, but in time the quartz itself is removed, leaving the great iron ore deposits of the Lake Superior region. This removal of the quartz is identical with that which can be measured in the weathering of a granite, where there is clear evidence of a solution of free quartz. In the case of the iron deposits, however, the removal is much more conspicuous for the reason that the silica is associated with a much more insoluble material and with none of the more soluble materials. Our attention is not diverted to the loss of more soluble materials. The process involves hydration, oxidation and carbonation with an increase in volume and liberation of energy. Fig. 8 summarizes

PLATE VIII



Representing losses and gains of constituents in the weathering of limesto
numbered f



and dolomite. Each broken line represents comparison of a pair of analyses, identification.

this alteration as determined by a large number of chemical, mineralogical, and volume measurements.

Hot waters may locally accomplish similar results. (See p. 137.)

AVERAGE CHEMICAL CHANGES DURING KATAMORPHISM OF SEDIMENTARY ROCKS

Some notion of the net chemical changes in sedimentary rocks during katamorphism is obtained by a study of the composition of river waters. Knowing somewhat definitely the relative proportions of substances yielded by weathering of igneous rocks to the surface solutions, and knowing the mineral contents of river waters, a study of the differences in these two groups of analyses should indicate roughly the effect of katamorphism of sediments in changing the relative proportions of substances in the river waters. This is discussed in Chapter II, Part III. The results accord very well with the inferences which might be drawn from the observed alterations of sediments. The most significant difference comes in the relative behavior of silica and the bases. It appears that the katamorphism of igneous rocks yields far more silica relative to bases than appears in the river waters, the explanation probably being that silica has reached the fairly stable condition of quartz in the sediments. As a complementary result the bases furnished by the katamorphism of sediments are relatively more important than they are in the katamorphism of igneous rocks. Especially is this true of lime, which in the sediments is in an easily soluble condition. Next in importance to the lime from the materials furnished by the sediments is soda, then magnesia, then potash.

CHAPTER VI

KATAMORHISM OF ORES

The katamorphism of acid and basic igneous rocks and sediments discussed on preceding pages includes the great mass of katamorphic alterations. The development of important iron ore and bauxite and clay deposits has already been described in this connection. Many other ores, especially sulphides, illustrate the processes of katamorphism. These ores are relatively insignificant in bulk as compared with the rocks discussed in preceding chapters, but being important commercially, have been intensively studied, with the result that a large amount of information is available in relation to their katamorphism. There is sometimes a tendency to discuss the alterations of ore bodies as unique. Recognition of the fact that the alteration of ores is merely a special case of katamorphism, controlled by certain general laws, adds to the effectiveness of study and interpretation of ore concentration. In turn the study of ores throws light on general katamorphic processes. Starting with a fresh igneous rock like a granite, katamorphism may produce end-products which are of little or no commercial value. From a syenite, from which free quartz is absent, the same process may, under favorable conditions, produce a bauxite, as in the case of the Arkansas bauxites. From ordinary basic igneous rock weathering may produce clays and carbonates of no commercial value, but exceptionally, as in the case of the Cuban serpentine rocks, exactly the same processes under favorable conditions have produced enormous bodies of iron ore. (See Chapter III, Pt. I.) Weathering of sedimentary formations poor in iron may yield clay and sand, but when the sediments are rich in iron, important iron ore deposits result. (See Chapter V, Pt. I.) It is obvious from these illustrations that the study of weathering as related to the concentration of the ore should not be separated from study of

katamorphism of rocks in general if the process is to be fully understood and interpreted.

Weathering of iron sulphide, which usually is an abundant constituent of sulphide ores, produces hydrated iron oxide, while iron sulphate, together with free sulphuric acid, is carried off in solution, to be precipitated in lower zones or ulitmately contributed to the run-off. The reaction follows:

$$2FeS_2 + 15O + 4H_2O = Fe_2O_3 + 4H_2SO_4.$$

The process is one of oxidation and hydration. Volume is increased when account is taken both of the residual material and that carried off. Energy is liberated. The sulphuric acid liberated by this reaction is an important agent in altering adjacent rocks and in altering ores further down. It performs work similar to that of carbon dioxide.

Zinc carbonate (smithsonite) or zinc silicate (calamine) is formed from the weathering of zinc sulphide, while a part of the zinc is removed in solution as zinc bicarbonate or zinc sulphate to be precipitated at other points. There is an increase of volume and liberation of energy.

From copper sulphides weathering produces copper oxides, copper carbonates and metallic copper. A part of the copper goes into solution as copper sulphate or carbonate to be carried off and precipitated elsewhere, part of it as chalcocite, enriching lower zones. The process is one of oxidation, hydration and carbonation.

Many other cases might be cited, but our purpose is merely to illustrate the importance of katamorphism in ore deposition.

Where there are silicates with the sulphides they are broken down to form clay, which remains in the oxide zone. Free quartz likewise largely remains, although it may become porous and spongy, due to slight leaching of the quartz as well as elimination of sulphides and carbonates. Commonly the sulphide deposits contain a variety of sulphide minerals, with calcite and quartz and silicates as gangue materials. In such combinations some are naturally altered more rapidly than others. Some form minerals more stable under weathering than do others. The result is a change in the vertical distribution of minerals. Frequently the upper few feet of the oxide zone may consist almost entirely of limonite, or other iron oxides, with clay and quartz, these sub-

stances being the most resistant to weathering. This constitutes the well known "iron capping." Farther down in the oxide zone, oxides and carbonates of copper, zinc, and other metals begin to appear, these having been leached from the higher zone. Clay often is more abundant than iron oxide in this part of the oxide zone, the suggestion being that in some cases it has been mechanically taken out from the iron capping or possibly dissolved by sulphuric acid. Still lower, usually below the water level at the time of concentration, are the secondarily deposited sulphides, leached from the upper zone.

The concentration of ores by katamorphic changes is effected in two ways: (1) residual concentration by the more rapid leaching of other constituents; (2) by addition of the metallic ores. In the oxide zone, the former method of concentration predominates; below the oxide zone both methods are common.

The story of concentration of metallic ores by katamorphic alterations is pretty well known qualitatively, but much remains to be learned when it is studied quantitatively. The changes in composition have not been quantitatively determined in enough cases to work out fully and definitely the differing rates of solution and removal and precipitation of the different substances under the different conditions. As the erosion plane works down each redeposited substance is again worked over.

The disseminated copper ores in porphyry afford an excellent example of downward secondary enrichment. In Fig. 9 the relations of copper content to depth in a single drill hole are represented graphically. Depth from surface is represented by the vertical scale at the right hand side of the diagram, and percentages of copper are platted on a horizontal scale shown at the base of the diagram. The primary sulphide zone at the bottom has an average copper content of 0.335 per cent. The secondary sulphide zone has an average copper content of 1.733 per cent. The oxide zone above the secondary sulphide zone has been pretty thoroughly leached of copper and the small part in which copper remains has an average copper content of 0.215 per cent. On the assumption that concentration has been entirely by downward enrichment, it has been calculated from the data in this diagram that the present secondary sulphide zone would require all of the copper from a primary sulphide zone containing an average

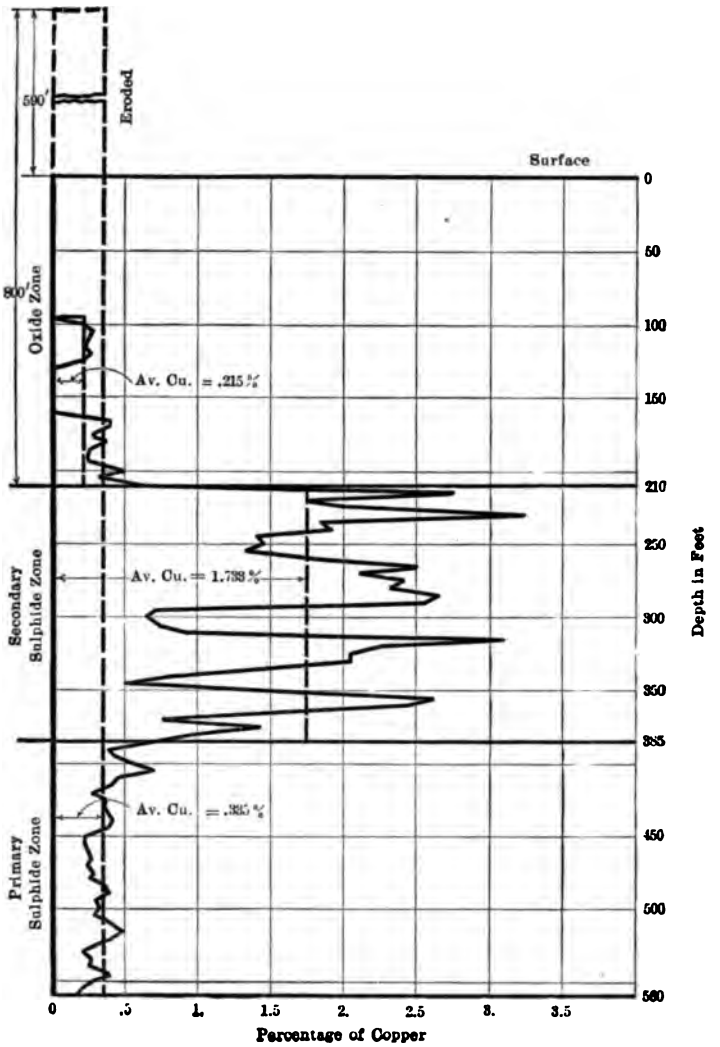


FIG. 9. Graphic representation of secondary concentration of disseminated copper ore in porphyry.

of 0.335 per cent copper 800 feet thick. This would mean the removal by erosion of 590 feet of material previously existing above the present surface. If a certain amount of copper were lost in the runoff during denudation, the required thickness would be greater than 590 feet. If concentration has been by lateral as well as downward movement of copper solutions, the thickness of material eroded would be greater where carried away from this particular locality, and less if copper were carried laterally toward this particular place. Obviously if the upward continuation of the original sulphide zone were richer than the present primary sulphide zone, the thickness of material eroded would be correspondingly less. The converse of this is also possible. It is evident that a complete study of the question would require similar study of a large number of drill holes together with investigation of the topographic history of the district and of the structural conditions controlling the downward or lateral flow of solutions. These data and the foregoing hypothetical conclusions are offered merely as a definite example of downward secondary enrichment and the possibilities of applying quantitative methods to a study of this question and no emphasis is placed on the specific results outlined, as more detailed study of the problem from a larger amount of data is plainly necessary before conclusions can be reached.

CHAPTER VII

THE REDISTRIBUTION OF THE CONSTITUENTS OF THE AVERAGE CRYSTALLINE AND IGNEOUS ROCK DURING KATAMORPHISM.

Preceding chapters have dealt with the katamorphic destruction of the igneous rocks, and, in a qualitative way, with the nature and distribution of resulting end-products. It has been pointed out that through the processes of weathering, erosion, transportation and deposition the constitutents of the igneous rocks are redistributed to form mainly the sedimentary rocks.

In this chapter katamorphic redistribution of the constituents of igneous rocks will be considered in terms of rocks, by weight and volume, in terms of minerals, and in terms of chemical constituents.

KATAMORPHIC REDISTRIBUTION OF AVERAGE IGNEOUS ROCKS IN TERMS OF SEDIMENTS (BY WEIGHT)

Sediments comprise three main classes, sands, clays, and carbonates. As most of the sediments within our range of observation are more or less cemented, it is more convenient to refer to them as sandstones, shales, and carbonate rocks. Attempts to determine the proportions of such sediments resulting from katamorphism of primary igneous rocks have been made by several methods:

(a) Figures for the total thickness and relative amounts of the three principal sedimentary rocks may be obtained from actual measurements, but such estimates are confined to the present continental areas, taking no account of suboceanic deposits, and limited to the areas which have been studied by geologists. Sections measured in different localities vary widely in the proportions of sediments. It is difficult to average sections properly because of irregularity in their distribution and secondary def-

ormation. Thus it is, that attempts to determine the relative proportions of sediments on the basis of actual measurement have given widely varying results, depending on the more or less arbitrary emphasis placed on certain sections or areas. The results accord, however, in indicating relatively larger proportions of sandstone and limestone, with reference to shale, than estimates based on comparison of composition of parent igneous rock and sediments, which are discussed below. An average of sections aggregating 520,000 feet, well distributed over North America, contains 46 per cent of shale, 32½ per cent of sandstone, and 21½ per cent of limestone. An average of sections aggregating 188,000 feet, scattered through Eurasia, gives 49 per cent of shales, 32½ per cent of sandstone, and 19 per cent of limestone. The proportions calculated below from average chemical composition on the following pages are 82 per cent shale, 12 per cent of sandstone, and 6 per cent of limestone. Reasons for this difference in proportions by the two methods of calculation are discussed on pages 67–68.

(b) Several estimates of the relative amounts of the principal sediments have been based on the nature and amount of material being carried to the sea by rivers, particularly on the ratio of calcareous to fragmental sediments. From a study of the nature and amounts of the dissolved mineral matter and solids carried by rivers, Reade [1] estimated that one-eighth to one-ninth of the transported materials are salts of calcium, and that limestone constitutes one-tenth of the sedimentary rocks. Van Hise,[2] in discussing Reade's estimates, calls attention to the fact that calcium carbonate constitutes 5 per cent of the shales and sandstones and that the more probable ratio of limestone to clastic sediments is 1 to 19.

(c) Some estimates have been made based either in part or entirely on average analyses of sedimentary and igneous rocks. Van Hise [3] combined average chemical analyses of the shales, sandstones, and limestones in the ratio of 65:30:5, respectively,

[1] Reade, T. Mellard, Chemical denudation in relation to geological time, p. 53.

[2] Van Hise, C. R., A treatise on metamorphism: Mon. 47, U. S. Geol. Survey, 1904, p. 941.

[3] Van Hise, C. R., A treatise on metamorphism: Mon. 47, U. S. Geol. Survey, 1904, p. 940.

and the resulting combination was found to approximate fairly well the composition of the average crystalline rock.

Clarke [1] estimated the average composition of the lithosphere to be 95 per cent igneous rock and 5 per cent sediments. He assigned 4.0 per cent to shales, 0.75 per cent to sandstones, and 0.25 per cent to limestones. This is equivalent to a ratio of shales to sandstone to limestone of 80:15:5, which accords fairly well with the following estimates by the writers. Clarke obtained his results by assigning all the free quartz of the average crystalline rock to the production of sandstone and half of the calcium of the average crystalline rock to the formation of limestone. The figure for free quartz was obtained from petrographical descriptions of about 700 igneous rocks.

Mead [2] estimated the proportions by weight of shale, sandstone, and carbonates to be 80:11:9 respectively. His method of calculation is more strictly quantitative in that it takes into account all of the principal constituents in comparing the composition of the parent igneous rocks with the composition of the several sedimentary products. This method is summarized below:

Redistribution of the original igneous and crystalline rocks may be expressed in the form of an equation as follows:

$$\text{Original crystalline rock} + \text{redistribution agencies} = \begin{cases} \text{Shales.} \\ \text{Sandstone.} \\ \text{Limestone and other carbonate rocks.} \\ \text{Vein filling materials.} \\ \text{Cementing materials.} \\ \text{Salts of the sea.} \\ \text{Residual matter.} \end{cases}$$

The consolidated sediments, in so far as they have undergone physical, chemical, and mineralogical changes during cementation (see pp. 104–105) are not "end-products" of katamorphism in the true sense of the term. The true end-products are the unconsolidated sediments, muds, clays, sands, and calcareous marls and oozes. However average chemical analyses of the unconsoli-

[1] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, p. 31.

[2] Mead, W. J., Redistribution of elements in the formation of sedimentary rocks: Jour. Geol., vol. 15, 1907, p. 250.

dated sediments are not available and it is necessary to employ the average analyses of the consolidated sediments and then to modify the results in accordance with the changes known to have taken place in the sediments during cementation.

If the several products making up the right-hand side of the equation could all be taken into account in proportion to their abundance, it is evident that the elements of the original rock would all be accounted for unless some minor factor has been overlooked. Of these end-products fair average chemical analyses are available for shales, sandstone, and limestone, and also for the salts of the sea. The sedimentary rocks form the major portion of the secondary material, the other products being only a small percentage of the total mass; hence some combination of the analyses of the three principal sediments should approximate the original rocks, though the omission of the minor constituents will cause certain discrepancies.

The problem, then, is to determine in what proportions the average analyses of shale, sandstone, and limestone should be combined to yield an average analysis as nearly like that of the average crystalline rock as possible.

The analyses used were those compiled by Clarke.[1] The problem was solved by a graphical method, described on pages 316–319.

Since the time of Mead's solution, Clarke has published revised and new average analyses of igneous rocks and sediments.[2] The average analysis of the igneous rocks represents "an average of all the analyses, partial or complete, made up to January 1, 1908, in the laboratories of the Survey." The average analysis of the shales represents an average of 78 rocks carefully selected and weighted. The average of the sandstones represents a composite analysis of 253 sandstones taken in equal weights. The average limestone analysis is a composite of 345 limestones. These analyses are given in Table VI.

[1] Clarke, F. W., Analyses of rocks from the laboratory of the U. S. Geol. Survey, 1880–1903: Bull. 228, U. S. Geol. Survey, 1904.

[2] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, pp. 24–28.

TABLE VI. AVERAGE ANALYSES OF IGNEOUS ROCKS AND SEDIMENTS

	Average igneous rock.	Average shale.	Average sandstone.	Average limestone.
SiO_2	60.17	58.90	78.64	5.20
Al_2O_3	15.05	15.63	4.77	.81
Fe_2O_3	2.59	4.07	1.08	.54
FeO	3.44	2.48	.30	
MgO	3.87	2.47	1.17	7.92
CaO	4.80	3.15	5.51	42.74
Na_2O	3.42	1.32	.45	.05
K_2O	3.00	3.28	1.32	.33
H_2O+	1.48	3.72	1.33	.56
TiO_2	.74	.66	.25	.06
P_2O_5	.26	.17	.08	.04
CO_2	.48	2.67	5.03	41.70
SO_3		.66	.07	.05
C		.82		
Misc.	.63			
	100.00	100.00	100.00	100.00

A solution of the problem with these new data indicates that the ratio by weight of the sediments, yielding a composite analysis most closely approximating the average igneous rock, is shale 81.7 per cent, sandstone 12.05 per cent, and limestone 6.25 per cent, or in even numbers shale 82 per cent, sandstone 12 per cent, and limestone 6 per cent.

The same process of solution has been applied to a series of five average analyses of crystalline rocks, ranging from acid to basic, and the same predominance of shale was observed in each case, ranging from 75 per cent for acid rocks to 88 per cent for the basic end of the series. This indicates that the redistribution of any crystalline rock results in a large predominance of shale in the resulting products.

In view of the fact that certain substances are actually added from the atmosphere and hydrosphere during katamorphism, it may be calculated that each 100 grams of igneous rocks yield by katamorphism 107.4 grams of sediments in proportions of 87.8 grams of average shale, 12.9 grams average sandstone and 6.7 grams average limestone.

If we take into account the salts of the sea as additional end-products of katamorphism, it may be calculated that 100 grams of average igneous rock yield 114 grams of end-products, consisting approximately of 87.8 grams of shale, 12.9 grams of sandstone, 6.7 grams of limestone, and 6.6 grams of ocean salts. The proportion of sediments to ocean salts may be approximated in the following manner: 100 grams of average igneous rock contain 3.47 grams of Na_2O. A weighted average of the sediments (Table XIII, p. 80) accounts for only 36 per cent of the sodium of the original igneous rock. If it is assumed that the sodium unaccounted for in the sediments (64 per cent of the sodium of the original rock) is contributed to the sea, this amounts to 2.25 grams of sodium for 100 grams of average igneous rock. From Dittmar's [1] analysis of the average salts of the ocean, the relative amounts of sodium, magnesium, calcium, and potassium, in terms of their oxides, are as follows:

Na_2O	81.7
MgO	12.4
CaO	3.3
K_2O	2.6
	100.0

If the sodium in the sea represents a loss of 2.25 grams of Na_2O, the total loss for the above oxides is 2.75 grams per 100 grams of original igneous rock. (2.25:81.7::x:100.) In terms of sodium chloride, magnesium sulphate, calcium carbonate, and potassium sulphate, this loss is equivalent to 6.6 grams of ocean salts per 100 grams of original rock.

The Composition of the Average Igneous Rock as Calculated from Analyses of Sediments

Examination of the discrepancies between the composition of the average igneous rock as determined by Clarke and a weighted average of the sediments raises the question as to whether Clarke's average of the igneous rocks actually represents the average of

[1] Van Hise, C. R., A treatise on metamorphism: Mon. 47, U. S. Geol. Survey, 1904, p. 942.

the igneous rocks of the earth's shell. Obviously, if the average of a large number of igneous rocks is to represent correctly the composition of the igneous rocks of the earth's crust it is necessary that the samples of the various types of igneous rocks be proportional to the amounts of these several types in the earth's crust.

It has been commonly assumed by geologists that the most wide-spread types of acid and basic rocks are granite and basalt, respectively. If this is true, it should be possible to closely approximate the average composition of all igneous rocks by some combination of the average granite and average basalt. If the average igneous rock may be represented by some combination of granite and basalt, it should be possible to determine what proportions of granite and basalt would yield an average which could be approximated most closely by some combination of shale, sandstone, and limestone.

This problem may be stated in the form of the following equation:

> x granite + y basalt = a shale + b sandstone + c limestone—given average analyses of granite, basalt, shale, sandstone, and limestone to solve for x, y, a, b, c.

Note that no assumption is made as to the relative abundance of shale, sandstone, and limestone. By a graphical method of solution [1] of the above problem, it has been found that a combination of 65 parts granite and 35 parts basalt best accords with the average composition of the sedimentary rocks. The average analyses of the granite and basalt employed were compiled by Daly [2] and are given in the following table:

[1] For description of method see Mead, W. J., The average igneous rocks: Jour. Geol., vol. 22, 1914, pp. 772–781.

[2] Daly, R. A., Average chemical compositions of igneous rock types: Proc. Am. Acad. Arts and Sci., vol. 45, 1910, p. 238.

TABLE VII. AVERAGE ANALYSES OF GRANITE AND BASALT

	1	2
SiO_2	70.47	49.65
Al_2O_3	14.90	16.13
Fe_2O_3	1.63	5.47
FeO	1.68	6.45
MgO	.98	6.14
CaO	2.17	9.07
Na_2O	3.31	3.24
K_2O	4.10	1.66
TiO_2	.39	1.41
P_2O_5	.24	.48
Totals	99.87	99.70
Fe	2.45	8.86

1. Average granite.
2. Average basalt.

For average analyses of the sediments see Table VI, p. 63. A comparison of the resulting combination of the granite and basalt analyses with Clarke's average of the igneous rocks is made in the following table:

TABLE VIII. COMPARISON OF AVERAGE IGNEOUS ROCK WITH COMBINATION OF AVERAGE GRANITES AND AVERAGE BASALT IN RATIO OF 65 TO 35

	1	2[1]	3	4
SiO_2	63.18	61.82	+2.36	+ 3.82
Al_2O_3	15.35	15.51	— .16	— 1.03
Fe_2O_3	2.97	2.67	+ .20	+ 7.5
FeO	3.35	3.45	— .10	— 2.9
MgO	2.79	4.02	—1.23	—30.5
CaO	4.58	4.96	— .38	— 8.45
Na_2O	3.28	3.51	— .23	— 6.5
K_2O	3.24	3.04	+ .20	+ 6.6

1. Average granite and average basalt, combined in ratio of 65 to 35.
2. Clarke's average igneous rock.
3. Differences between 1 and 2, on basis of total rock.
4. Differences between 1 and 2, in percentage of each constituent.

[1] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, p. 25.

The differences between columns 1 and 2 are small. The most striking result of the comparison is the marked similarity between the combined granite and basalt and Clarke's average analysis. The greatest differences are for silica and magnesia, the combination of granite and basalt being slightly more acidic than the other averages. The greatest difference is in MgO, the combined granite and basalt showing 30.5 per cent less MgO than Clarke's average. This difference is so much greater than for any of the other oxides that it may be of significance. It is possible that igneous rocks collected largely for petrographic purposes would include the unusual and interesting alkaline and ultra-basic types, the syenites, peridotites, etc., in an amount out of proportion to their abundance as compared with the more common granite and basalt types. The alkalies, lime, silica, and alumina would not be materially affected by this, as the alkaline rocks would offset the basic varieties, but the range of MgO content being much greater (varying from 1 per cent or less in the alkaline rocks to 30–40 per cent in the peridotites), this oxide would tend to be increased by the inclusion of an excess of the less common rocks.

Comparison of Calculated Proportions of Sediments with those Observed in Geologic Sections

The nature of the data available for the above solutions precludes the possibility of closely accurate results, but the proportion of sediments named satisfies the requirements in regard to all the constituents so closely that we believe the figures can be accepted with some confidence. Any other combination of sediments would give considerable excess or deficiency of certain constituents in the sediments, difficult to explain. The substantial accuracy of the proportions named, and especially the great predominance of shale, is further indicated by the fact that they do not differ in this essential feature from several estimates based on chemical and mineralogic data previously made by others.

The most marked differences are with the estimates made by measuring actual sections. As already indicated, such measurements cannot be expected to yield accurate results, but they uniformly show higher percentage of sandstone or limestone than the proportions here calculated by analysis. This difference presuma-

bly finds its explanation in the fact that in the sediments as they exist, there are many in which there has not been a complete separation into the typical sands, shales, and carbonates, represented by the average analyses here used. The sands, for instance, include many terrestrial and subaërial formations of enormous thickness, which have resulted mainly from disintegration, without separation into constituent carbonate, clay and sand elements. Also the sections disclose a marked abundance of shaly limestone without correlative abundance of calcareous shales. These are probably not represented adequately in the average analyses. The proportions of sediments calculated from average analyses of type sediments are really the proportions there would be if all of the sediments reached the ultimate form represented by these analyses, which they obviously do not. Yet it is not at all certain that these considerations fully explain the higher percentages of sandstone and limestone in the actual sections. There is a possibility that during repeated cycles of metamorphism there has been a progressive increase in proportions of sandstone and limestone due to separation of quartz and carbonates from the shales, and by certain other cumulative effects of metamorphic processes discussed on pages 271–273.

Underlying all of the following quantitative discussion is the assumption that sediments develop from the katamorphism of average igneous rocks in the approximate proportions indicated above. In so far as sediments are mixed and are not represented by the average analyses here used (p. 12) the actually existing proportions of sediments may be different, although the totals of elements in the sediments as a whole will not differ. It is believed that the proportions of sediments calculated from the average analyses of typical materials are nearly enough correct to afford a basis for a quantitative discussion of the distribution of the elements and minerals among the sediments as a group. Discussion based on this assumption brings out some interesting points, which seem to be established even if the proportions used are not exact. Too much stress should not be laid upon the absolute figures quoted, but attention should be given rather to the relative distribution in the different classes of sediments.

KATAMORPHIC REDISTRIBUTION OF AVERAGE IGNEOUS ROCKS IN TERMS OF SEDIMENTS (BY VOLUME)

The products of katamorphic redistribution of the average igneous rock occupy a greater volume than the original rock, due to: (a) addition of materials, principally water, carbon dioxide, sulphur trioxide, chlorine, and oxygen, from the hydrosphere and atmosphere; (b) development of minerals of lower specific gravity; and (c) development of porosity.

(a) 100 grams of average igneous rock contribute approximately 97 grams to the average sediments, the remaining 3 grams going to the salts of the sea. (See p. 64.) Including water, carbon dioxide, oxygen, sulphur trioxide, and other constituents of the sediments, derived from the hydrosphere and atmosphere, the total weight of sediments from 100 grams of igneous rock is 107.4 grams. (See p. 63 and Table IX.) If the specific gravity of the sediments did not differ from that of the average igneous rock, the increase in volume, due to addition of material, would therefore amount to 7.4 per cent.

(b) The average specific gravity of the sediments, excluding pore space, weighted in proportion to their volumes, is 2.71. The average specific gravity of the igneous rock is 2.8, consequently 100 grams of igneous rock occupy 35.67 cubic centimeters and 107.4 grams of sediments occupy 39.6 cubic centimeters; therefore, 100 units of volume of igneous rocks yield 111 units of volume of sediments. Considering only the addition of material and the lower average specific gravity of new minerals, and neglecting pore space, the increase in volume is 11 per cent.

(c) Little information is available as to the porosity of shale. Sorby [1] determined the average porosity of certain English coal measure shales to be 13.2 per cent. An average of 76 determinations of porosity of American and English sandstones is 14 per cent. The fact that these rocks were used for building purposes indicates that the average determined is probably lower than the actual average of the sandstones as a whole, as the rocks desirable for structural work are those of low porosity. An average of 53 determinations of porosity of American and English limestones

[1] Sorby, H. C., Application of quantitative methods to the study of the structure and history of rocks: Quar. Jour. Geol. Soc. London, vol. 64, 1908, p. 229.

is 11.5 per cent. These represent determinations made on limestone used for structural purposes and consequently the average is probably lower than the average porosity of the carbonate rocks. The limestones and dolomites are characteristically cavernous and these openings obviously would not appear in determinations of porosity in specimens. The average figures employed for porosity therefore represent something less than the true average porosity of the consolidated sediments. The figures available show an average porosity for sediments of 13 per cent which represents the increase of volume due to development of voids.

The increase in volume due to all the factors above mentioned, addition of material, development of minerals of low specific gravity, and development of porosity, amounts to at least 28 per cent.

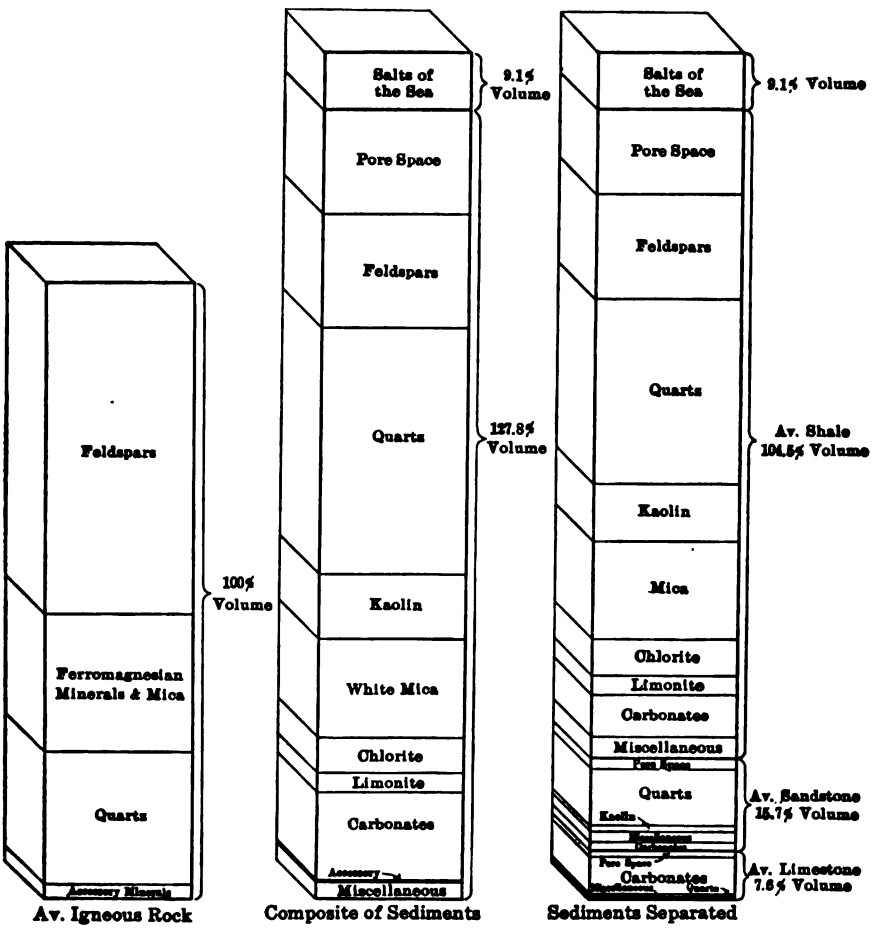
It has been calculated (see p. 64) that sodium, magnesium, calcium, and potassium are contributed to the sea to the amount of 2.75 grams of their oxides per 100 grams of average igneous rock. If we calculate this material in terms of sodium chloride, magnesium chloride, calcium carbonate, and potassium sulphate, the total weight of these salts is 6.58 grams. With an average specific gravity of 2.01 their volume is 3.27 cubic centimeters. If this volume is included with the volume of the sediments the volume increase becomes 36.9 per cent.

The data and calculations employed in arriving at these volume changes appear in Table IX.

The mineral composition of the average igneous rock and of the average sediments expressed in terms of volume are given in Table X.

The volume changes described above are expressed graphically in Plate IX.

PLATE IX



Representing volume changes in the production of sandstone, shale, and limestone from average igneous rocks by katamorphism.

TABLE IX. SHOWING PROPORTIONS OF SEDIMENTS AND OCEAN SALTS IN TERMS OF VOLUME AND WEIGHT

	A	B	C	D	E	F	G	H	I	Volume of ocean salts			
Consolidated sedimentary rocks.	Proportions, considering only constituents derived from igneous rocks.	$\frac{97}{100}$ x A = B Grams of "derived" constituents in sediments per 100 grams average igneous rock.	100-(CO_2 + H_2O etc.) Per cent of "derived" constituents in sediments.	$\frac{B}{C}$ x 100= D Grams per 100 grams average igneous rock including "derived" & "underived" constituents.	Average mineral specific gravity.	$\frac{D}{E}$ = F. Cubic centimeters per 100 grams average igneous rock.	Volume of sediments (excluding pore space) per 100 units of volume of average igneous rock.	Average porosity of sediments.	$\frac{G}{100-H}$ x 100=I. Volume of sediments including pore space per 100 units of volume of average igneous rock.	Grams per 100 grams average igneous rock.	CC per 100 grams average igneous rock.	Per cent of volume of average igneous rock.	Net volume increase including ocean salts.
Shale.....	83.5	81.0	92.17	87.8	2.71	32.4	90.8	13%	104.5	NaCl 4.18	1.94	5.4	
Sandstone:....	12.5	12.1	93.56	12.9	2.67	4.8	13.5	14%	15.7	$MgSO_4$ 2.09	1.19	3.3	
										$CaCO_3$ .18	.07	.2	
Limestone	4.	3.9	57.86	6.7	2.76	2.4	6.7	11.5%	7.6	K_2SO_4 .13	.07	.1	
Totals....	100.0	97.0		107.4	Av. 2.71	39.6	111.0		127.8	6.58	3.27	9.1	136.9
Unconsolidated sediments													
Clay and mud.	83.5	81.0	91.80	88.3	2.68	32.9	92.2	27.%	126.				
Sand.....	12.5	12.1	93.56	12.9	2.67	4.8	13.5	30.%	19.3				
Marls, oozes, etc.....	4.	3.9	57.86	6.7	2.76	2.4	6.7	25.%	8.9				
Totals....	100.0	97.0		107.9		40.1	112.4		154.2				

Volume Change in Development of Unconsolidated Sediments

The above figures for increase of volume are based on comparison of igneous rocks and the average sands, shales, and limestones. If the unconsolidated equivalents of these sediments be considered the volume increase is still larger.

By recasting the average shale analysis, including sufficient water to express the alumina as kaolin (in which form it largely occurs in unconsolidated sediments) rather than mica (the common form in consolidated sediments), the total weight of unconsolidated sediments per 100 grams of average igneous rock becomes 107.9. This reduces the average specific gravity of the sediments to 2.69, and the volume of the sediments, exclusive of pore space, produced from 100 units of volume of average igneous rock is 112.4. The average porosity of unconsolidated clays and muds varies between wide limits, being as low as that of shale in some cases, and as high as 70 to 80 per cent in certain water deposited muds and some residual clays. Sorby [1] determined the porosity of a number of English clays, the average of which is 26.7. An average of 19 American clays is 26.38.[2] A figure of 30 per cent for the average porosity of unconsolidated sand has been used. The porosity of the sands does not vary between very wide limits and this figure is in accord with the theoretical and determined porosities. Little data are available on the porosity of unconsolidated, calcareous sediments. Sorby [3] has investigated the subject and the figure of 25 per cent employed in Table IX is in accord with his results.

Assigning an average porosity of 27 per cent to the clays and mud, 30 per cent to the sands and 25 per cent to the unconsolidated calcareous sediments, their total volume, including pore space, is 154 per cent of the volume of average igneous rock, an increase of 54 per cent.

[1] Sorby, H. C., Application of quantitative methods to the study of the structure and history of rocks: Quart. Jour. Geol. Soc. London, vol. 64, 1908, p. 229.

[2] Ries, H., Clays, their occurrence, properties, and uses, New York, 1908, pp. 163 and 165.

[3] Sorby, H. C., Application of quantitative methods to the study of the structure and history of rocks: Quart. Jour. Geol. Soc. London, vol. 64, 1908, p. 229.

Total Volume of Sediments Calculated From Amount of Ocean Salts

It has been calculated (p. 64) that 100 grams of average igneous rock contribute 2.75 grams of material in terms of oxides to the dissolved mineral matter of the ocean. The total mineral matter of the ocean [1] has been calculated by Clarke from Dittmar's analyses, with Karstens' value for the volume of the ocean, to be 46,188.0 x 10^{12} metric tons. From the data from which this total is derived the amount of sodium, potassium, calcium, and magnesium *in terms of oxides* is 23,252 x 10^{12} metric tons. If this represents 2.75 per cent of the total amount of igneous rocks required to supply the ocean salts and sediments, this amount is 847,000 x 10^{12} metric tons. With an average specific gravity of 2.8 this amounts to 72,000,000 cubic miles of igneous rock.

The change from igneous rocks to sedimentary rocks involves an increase in volume of 28 per cent (Table IX, p. 71) hence the volume of sediments derived from this amount of igneous rock would be 92,000,000 cubic miles.

The superficial area of the earth is approximately 200,000,000 square miles, hence this volume of sediments is equivalent to a layer 0.46 mile thick enveloping the earth. The continental platforms constitute approximately one-third of the area of the earth or 66,000,000 square miles. The above volume of sediments would account for an average thickness of 1.39 miles over this area. In terms of the principal classes of sediments this would be equivalent to an average thickness of 6,000 feet of shale, 900 feet of sandstone, and 500 feet of limestone.

On the basis of the ratio of sodium contributed to the sediments and to the ocean from the average igneous rocks Clarke [2] has estimated the volume of sediments to be equivalent to a shell enveloping the earth with an average thickness of between 0.4 and .5 of a mile.

[1] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, p. 125.

[2] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, pp. 29–30.

KATAMORPHIC REDISTRIBUTION OF IGNEOUS ROCKS IN TERMS OF MINERALS (BY WEIGHT)

Average Mineral Composition of Igneous Rocks

The average igneous rock is essentially equivalent in chemical composition to a combination of an average granite and an average basalt in proportions of 65 to 35. (See pp. 64–67.) Since granite and basalt are the dominant types of acid and basic rocks, it seems reasonable to assume that a combination of the average mineral compositions of these two types in the above proportions would express fairly well the average mineral composition of the igneous rocks. In Table X are given the mineral composition

TABLE X. MINERAL COMPOSITIONS OF AVERAGE GRANITE AND AVERAGE BASALT CALCULATED FROM AVERAGE ANALYSES IN TABLE VII, P. 66

Average granite	Average basalt	Average igneous rock Granite = 65% Basalt = 35%
Albite 26.10	Albite 24.70	Albite 25.60
Anorthite 9.30	Anorthite 10.70	Anorthite 9.80
Orthoclase 17.05	Orthoclase 10.75	Orthoclase 14.85
Biotite 5.94	Augite 36.90	Biotite 3.86
Muscovite 5.93	Olivine 7.58	Muscovite 3.85
Hornblende 2.46	Magnetite 5.80	Hornblende 1.60
Quartz 31.39	Ilmenite .73	Augite 12.90
Magnetite 1.72	Titanite 2.84	Olivine 2.65
Ilmenite .31		Quartz 20.40
		Magnetite 3.15
		Titanite and ilmenite 1.45
Totals 100.20	100.00	100.11
Summary:		
Feldspars 52.45	46.15	50.20
Ferro-magnesian minerals & mica 14.33	44.48	24.80
Quartz 31.39		20.40
Accessory minerals 2.03	9.37	4.60
Totals 100.20	100.00	100.00

of the average granite and of the average basalt, and of a composite of these two rocks in the proportion of 65 to 35. These mineral compositions express as nearly as possible the actual mineral composition of the rocks. In other words, the mode rather than the norm has been calculated. The average analyses used for the granite and basalt, compiled by Daly, are given in Table VII, p. 66.

Clarke has made the following estimate of the mean mineralogical composition of the igneous rocks from a statistical examination of about 700 igneous rocks, which have been described petrographically.

TABLE XI. CLARKE'S ESTIMATE OF AVERAGE MINERALS OF IGNEOUS ROCKS [1]

Quartz	12.0
Feldspars	59.5
Hornblende and pyroxene	16.8
Mica	3.8
Accessory minerals	7.9
Total	100.0

The essential difference between Clarke's estimate and the estimate given above is in the quartz and feldspar percentages. If granite makes up 65 per cent of the igneous rocks of the earth's crust, Clarke's estimate of 12 per cent quartz in the average igneous rock implies an average quartz content for the granites of only 18.45 per cent, which is obviously too low. The average percentage of quartz in 84 norms,[2] calculated from as many analyses of plutonic granites by Iddings, is 32.80. Since normative quartz is in general less than the actual percentage, the quartz content as calculated for the average granite, Table X, is apparently not too high.

[1] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, p. 31.

[2] Iddings, J. P., Igneous rocks, vol. 2, 1913, pp. 114–146.

Average Mineral Composition of Sediments

Average mineral compositions for the shales, sandstones, and limestones, calculated from the average chemical analyses in Table VI, are given in the following table:

TABLE XII. AVERAGE MINERAL COMPOSITIONS OF SEDIMENTS CALCULATED FROM AVERAGE ANALYSES IN TABLE VI, P. 63

	Average shale	Average sandstone	Average limestone	Combined sediments. Shale 82% Sandstone 12% Limestone 6%
Quartz	31.91	69.76	3.71	34.80
Kaolin	10.00	7.98	1.03	9.22
White mica	18.40	.....	.....	15.11
Chlorite	6.40	1.15	.....	5.29
Limonite	4.75	.80	.....	4.00
Dolomite	7.90	3.44	36.25 [1]	9.07
Calcite	.....	7.21	56.56	4.25
Gypsum	1.17	.12	.10	.97
Orthoclase	12.05	8.41	2.20	11.02
Albite	5.55	.....	.....	4.55
Magnetite	.....	.58	.....	.07
Rutile	.66	.12	.06	.55
Ilmenite	.....	.25	.....	.02
Apatite	.40	.18	.09	.35
Carbon	.81	.....	.....	.73
Total	100.00	100.00	100.00	100.00

[1] Includes small amount of $FeCO_3$.

For further discussion of the minerals of the sediments see Chapters I, II, and III, Part II.

Redistribution of Quartz

The average percentage of free quartz in the igneous rocks has been estimated (Table X) as 20.4 per cent, and in a composite of the sediments as 34.80 per cent (Table XII). Since 100 grams of average igneous rock yield 107.4 grams by weight of sediments (see p. 63) the amount of quartz in the sediments comparable to the quartz of the igneous rock is 34.8 per cent of 107.4 grams, or

37.4 grams. This increase in quartz from 20.4 per cent in the igneous rocks to 37.4 per cent in the sediments is due to the liberation of silica by the katamorphism of the silicates of the igneous rocks. Much silica is freed by the breaking down of the feldspars, particularly of the orthoclase and albite molecules.

As a greater part of the silica goes into solution upon the destruction of silicates and is deposited as some of the forms of the quartz, it follows that the amount of free quartz is greatly increased during repeated cycles of katamorphism and erosion.

The amount of silica which would be liberated by complete katamorphism of the average igneous rock may be estimated approximately by assuming that, after katamorphism, alumina combines with silica in the proportion of $Al_2O_3.2SiO_2$, and by subtracting the amount of original free quartz in the igneous rock from the silica which remains after making this combination. The average igneous rock contains 15.51 per cent Al_2O_3, which requires 18.25 per cent SiO_2 to form $Al_2O_3.2SiO_2$. The total silica is 61.82 per cent. Subtracting from this the 18.25 per cent assumed to be combined with alumina, and the 20.47 per cent free quartz of the igneous rocks, there remains 23.17 per cent free silica which may be supposed to represent the amount freed in a case of complete katamorphism where silica remains in combination with alumina in ratio of two to one. The silica thus freed, together with the original quartz, amounts to a total of 43.57 per cent of free quartz. This figure exceeds by only 6.3 per cent (43.7—37.4=6.3) the amount of free quartz in a weighted average of the sediments, indicating that the sediments represented by the average analyses are as a group katamorphosed nearly to the above assumed limit.

The 37.4 grams of quartz in the composite of the sediments are distributed as follows: shale 28 grams, sandstone 9 grams, limestone 25 grams. Of the 20.4 grams of quartz in the original rock only 9 grams are accounted for in the sandstone, which leaves 11.4 grams for the shale.

Redistribution of the Feldspars

So far as the feldspars of the original rock escape katamorphism, they are transported and deposited as feldspars, the coarser grains with the sand and the finer particles with the muds and clays. Fragments of mechanically transported feldspars are found es-

pecially in the sands, derived from rapid erosion of igneous rock areas, as in certain arid regions of high relief. The feldspar of the sedimentary rocks, which amounts to about 12 per cent, represents mainly undecomposed feldspars from igneous rocks, schists, and gneisses. It is possible that a part of the feldspar of the sediments is secondary and has been developed since the deposition of the sediments by recrystallization. (See pp. 115, 130–131.) Feldspars which are decomposed by katamorphism before transportation and redeposition yield a portion of their bases in the form of soluble salts, such as the carbonates, chlorides and sulphates. Part of their silica content is liberated and removed in solution. The alumina, with the remainder of the silica, and part of the bases, particularly soda and potash, become hydrous aluminum silicates of the kaolin group, or white mica, which are transported mechanically and contributed to the finer fragmental sediments.

Redistribution of Ferromagnesian minerals

So far as these minerals escape katamorphism during redistribution, they may be transported and deposited with the fragmental sediments in an unaltered condition. Clays and sands derived from areas in which crystalline rocks are abundant, particularly the basic varieties, contain more or less unaltered ferromagnesian fragments, the amounts depending upon the relative rates of decomposition and erosion on the land surface. In general ferromagnesian minerals are more largely destroyed by katamorphism than the feldspars. The reason for this is probably their basic character which renders them more susceptible to attack by the acid agencies of katamorphism. As intermediate stages in their destruction, hornblende may form from augite, chlorite or talc or serpentine from augite or hornblende, or other ferromagnesian products may form. These minerals, especially chlorite and serpentine or talc, may occur in small amounts in sediments. For the most part, however, the ferromagnesian minerals are destroyed and their constituents appear in the sediments as simple oxides, carbonates, or hydrates. As soon as sediments begin to be anamorphosed, either by cementation or by dynamic or contact metamorphism, the ferromagnesian minerals begin to reappear.

Redistribution of Principal Accessory Minerals

Magnetite is an essential rather than an accessory mineral in some basic rocks, but for the igneous rocks as a whole probably should be classed as accessory. This mineral is resistant to alteration and is probably in large part contributed to the mechanical sediments in an unaltered form. Its form, density, and hardness cause it to be largely segregated with the sands. Oxidation and hydration change magnetite to limonite and hematite.

Ilmenite, like magnetite, resists alteration and is largely redistributed mechanically.

Titanite is in part redistributed mechanically but also alters to ilmenite, leucoxene, and rutile. It probably accounts for much of the rutile of the shales and slates. (See p. 107.)

Zircon is extremely resistant to alteration and solution. Its hardness and density cause it to be largely segregated with the sands.

Apatite is comparatively stable and is redistributed mechanically to a considerable extent. It is altered to soluble phosphates by acids and is a source of phosphoric acid for plants.

KATAMORPHIC REDISTRIBUTION OF IGNEOUS ROCKS IN TERMS OF CHEMICAL CONSTITUENTS (BY WEIGHT)

In the following table the average analyses of the sedimentary rocks in Table VI have been combined in the proportions stated above, and the resulting combination compared with the average igneous rock.

TABLE XIII. COMPARISON OF COMBINATION OF SEDIMENTS WITH AVERAGE IGNEOUS ROCK

	Grams	SiO_2	Al_2O_3	Fe_2O_3	FeO	Fe	MgO	CaO	Na_2O	K_2O	H_2O+	TiO_2	P_2O_5	CO_2	SO_3	C
Average shale....	87.8	51.70	13.71	3.57	2.18	4.20	2.17	2.76	1.16	2.88	3.26	.58	.15	2.34	.58	.72
Average sandstone	12.9	10.150	.615	.139	.039	.130	.151	.711	.058	.170	.172	.032	.001	.649		
Average limestone	6.7	.348	.054	.018	.018	.026	.530	2.860	.003	.022	.038	.004	.002	2.793	.003	
Total sediments.	107.4	62.198	14.379	3.727	2.237	4.356	2.851	6.331	1.221	3.072	3.470	.616	.153	5.782	.583	.72
Average igneous rock.....	100.00	60.20	15.05	2.59	3.44	4.49	3.87	4.80	3.42	3.01	1.48	.74	.26	.48	.28[1]	
Differences in grams per 100 grams average igneous rock		+1.998	—.671	+1.137	—1.203	—.14	—1.019	+1.531	—2.199	+.062	+1.99	—.124	—.107	+5.302	+.303	
Difference in terms of per cent of amount of each constituent in average igneous rock		+3.32	—1.46	+43.9	—35.0	—3.10	—26.3	+31.9	—64.2	+2.06	+130.0	—16.7	—41.0	+1104.0	+108.0	

[1] Calculated from 0.11 S in average igneous rock.

Again it should be remembered that underlying all of the following quantitative discussion is the assumption that sediments develop in the approximate proportions indicated above from the katamorphism of average igneous rocks.

Constituents added From the Hydrosphere and Atmosphere During Katamorphism of Igneous Rocks

To each 100 grams of igneous rock have been added during katamorphism to sediments 5.30 grams of CO_2, 1.99 grams of H_2O, 0.30 gram SO_3, .72 gram carbon, and approximately 1. gram of oxygen, amounting to a total of 9.36 grams of material from the atmosphere and hydrosphere. Since the shales have suffered some dehydration during cementation, the amount of water added during katamorphism is more than 1.99 grams. If the average shale be calculated as converted into clay by addition of combined water, the amount of water added during katamorphism would be 4.0 grams.

Redistribution of Silica

From the data in Table XIII, page 80, it appears that 83 per cent of the silica of the sedimentary rocks is in the shales, 16.5 per cent in the sandstones, and about 5 per cent in the limestones. The silica content of the combined sediments exceeds by 3.32 per cent the silica available in the original rocks. This discrepancy is relatively small and probably within the limit of error of the data employed.

During the processes of redistribution silica is transported mechanically and in solution. Silica is carried in solution as colloidal silicic acid and in the form of soluble silicates, particularly the silicates of the alkalies. It may be deposited in the openings of the rocks as quartz or chalcedony. Silica which gets into the "runoff," that is into the rivers and streams, is ultimately deposited either by chemical or organic agencies as various forms of quartz, and to a less extent as silicates. A large part of the silica carried in solution by rivers is deposited with the muds and clays. Some silica is deposited as beds of chert in carbonate and shale formations.

Silica carried mechanically in the form of quartz grains may be sorted by winds and running water and deposited as sand. Many

of the finer particles of quartz are not separated from the clays and muds and are deposited with them in the great shale formations. A large part of the silica transported mechanically is in combination with alumina and water, forming a considerable portion of the clay minerals. This silica is represented largely in the clays and shales. The silica, transported in the form of undecomposed silicates of the crystalline and igneous rocks, is ultimately deposited as part of the mechanical sediments, the sand and clay deposits.

The relative amounts of silica transported mechanically and in solution obviously depend on the relative speed of decomposition and erosion. Decomposition first liberates silica from the silicates, but ultimately the free quartz is attacked. When erosion lags sufficiently behind decomposition practically all of the minerals containing silica yield silica to solution. Under certain extreme conditions of weathering practically all of the silica of a rock, even the silica of the clays, may be removed in solution. This is illustrated by the Cuban iron ores and the Arkansas bauxite deposits. (See Chapter III, Part I.) Quartz itself is more or less soluble under weathering conditions, as illustrated in the weathering of a granite from Georgia (Pl. III), where the quartz in the altered rock is considerably less than that originally present in the fresh rock. The solution of quartz is also illustrated in the concentration of the Lake Superior iron ores, which have been formed from a rock made up of iron oxide and finely crystalline quartz, by the removal of quartz at the rate of over a ton of silica per ton of ore.

Under conditions of rapid erosion, where decomposition has less opportunity to act, the proportion of silica removed in solution is less, and more of the silica is carried mechanically in the form of quartz and disintegrated particles of the original silicates.

Redistribution of Alumina

From the data in Table XIII, page 80, alumina is distributed among the sedimentary rocks in the following proportions: shale 95.3 per cent of total alumina in sediments, sandstone 4.3 per cent, limestone 0.4 per cent.

In the weighted average of the sediments there is a deficiency amounting to 4.46 per cent of the alumina in the original rocks.

This discrepancy is small and is probably within the limit of error due to inaccuracies in the average analyses used. In so far as high alumina clays and bauxites in the residual rock mantle have been omitted from consideration, this would tend to account for the deficiency of alumina in the sediments.

Weathering yields very little alumina to solution and consequently the transportation of alumina during katamorphism is almost entirely mechanical. It is carried in the form of hydrous aluminum silicates, such as kaolin and the related clay minerals, some of them probably in colloidal form, as hydrous aluminum silicates of the several bases, such as sericite, chlorite, etc., and also to some extent as unaltered remnants of the original aluminum-bearing silicates of the igneous and crystalline rocks. To a very small extent it is carried as hydrous aluminum oxides. Practically all of the alumina is contributed to the mechanical sediments and it is probable that the alumina in the limestone represents fragmental material.

Redistribution of Iron

The iron of the sedimentary rocks is present in the three principal sediments in the following ratios: Shale 96.5 per cent of the total iron in sediments, sandstone 2.9 per cent, limestone 0.6 per cent. (See Table XIII, p. 80.)

A weighted combination of the sedimentary rocks accounts for all but 3.1 per cent of the iron in the original parent rocks. In so far as this difference is a real one it may be accounted for at least in part by the segregations of iron in iron formations and iron deposits which have not been considered in the average analyses of the sediments. On the other hand, some iron may have been contributed to the sedimentary rocks by primary solutions from magmatic sources. The ferric salts of iron are insoluble as such, and are only taken into solution under reducing conditions which change them to the ferrous form. In this form iron is carried principally as the carbonates, sulphates, or humates. In the weathering of rocks, both sedimentary and igneous, the amount of iron taken into solution depends upon the nature of the solutions acting. If the conditions favor oxidation, the iron tends to remain in an insoluble form, to be ultimately removed mechanically with the clay. If the conditions are not oxidizing,

the iron may be taken into solution in the ferrous form and carried until oxidized or otherwise precipitated.

Deposits of iron ore are formed as an incident in the redistribution of iron by katamorphism, principally by the following processes: (a) by the weathering of igneous or sedimentary rocks low in alumina in which the iron remains as insoluble ferric oxide, while the other more soluble constituents are removed in solution. This insoluble residue of ferric oxide, if sufficiently free from impurities, constitutes an iron ore. In this class are the iron ores of the Lake Superior region, the residual brown ores commonly developed from limestones, the residual lateritic ores of the Cuban type, and many other deposits of types similar to the above; (b) by the transportation of iron in solution and its deposition in a state of sufficiently pure iron mineral to yield an iron ore. In this class are the bog iron ores, probably in part the Clinton granular ores, and the Brazilian hematites: (c) by transportation mechanically with possibly natural sorting and concentration of previously existing residual bodies of iron ore.

Redistribution of Magnesia

Of the total magnesia in the sedimentary rocks 76 per cent occurs in the shales, 18.7 per cent in the limestone, and 5.3 per cent in the sandstone.

Magnesia is carried in solution principally as carbonates and sulphates. Magnesia enters into the composition of several minerals which are comparatively stable under surface conditions and remain with the residuals of weathering; consequently this oxide is carried mechanically to a considerable extent. The principal minerals of this type are chlorite, talc, and other hydrous magnesian silicates, with or without alumina, of a similar nature. These minerals are carried and deposited with the clay minerals as muds and clays, ultimately entering into the composition of the shales. The magnesia carried in solution is, during its underground passage, deposited as cementing and vein-filling material in the form of carbonates and that part which escapes to the sea is largely deposited as dolomite, though some of it remains in solution in the ocean as chloride and sulphate.

A weighted combination of the sediments (Table XIII, p. 80) fails to account for the magnesia in the average igneous and crys-

talline rocks by 26 per cent of the magnesia in the igneous rocks. This difference is a large one and requires consideration. The lime present in the sediments (Table XIII, p. 80) exceeds by 32 per cent the lime in the original rocks. This deficiency of magnesia and excess of lime suggest the possibility that the average analysis of the limestones used (Table VI, p. 63) gives more lime and less magnesia than the true average of the limestones. The average limestone has 42.74 per cent CaO and 7.92 per cent MgO. If we change these percentages to those of a corresponding dolomite, they become 28.2 per cent CaO and 20.15 per cent MgO. Using these figures in computing the average lime and magnesia contents of the sediments we find a deficiency of 5.2 per cent of the magnesia of the parent rocks and an excess of 11.7 per cent of the lime available in the parent rocks. Thus by taking a limiting case (certainly the average limestone is not a dolomite), we fail to account for all of the discrepancies of lime and magnesia on this basis.

It is possible that part of the deficiency of magnesia in the sediments is in the large amount of that constituent in solution in the sea. Clarke [1] computes from Dittmar's estimates of the salts of the sea that the sea water contains $1,721 \times 10^{12}$ metric tons of magnesium, equivalent to $2,865 \times 10^{12}$ metric tons of magnesia. The total volume of sediments has been computed to be (p. 73) 92,000,000 cubic miles equivalent to $910,000 \times 10^{12}$ metric tons. If the average magnesia content is 2.62 per cent ($.82 \times 2.68 + .12 \times 1.25 + .06 \times 13.74$), this mass of sediments contains $23,800 \times 10^{12}$ metric tons of MgO. The ratio of the magnesia in the sediments to magnesia in the sea is then 23,800: 1,721 or the magnesia in the sea is 6.75 per cent of the total. As the discrepancy of magnesia between igneous rock and sediments to be accounted for is 26 per cent of the total, this leaves a deficiency of over 19 per cent of the magnesia still to be accounted for.

Still another possible explanation is suggested by the fact that the average igneous rock may contain an undue proportion of rocks high in magnesia because of their petrographic prominence. (See pp. 64–67.) This would give an average high in magnesia

[1] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, p. 125.

and would explain at least in part the apparent deficiency of magnesia in the sediments.

Redistribution of Lime

The lime of the sedimentary rocks is distributed among the principal classes of sediments in the following proportions; shale 43.6 per cent, limestone 45.2 per cent, sandstone 11.2 per cent.

This oxide is one of the most soluble of the rock constituents, and where erosion lags behind decomposition, practically all of this constituent is removed in solution. In the natural waters, lime is carried principally as the bicarbonate. Lime in the form of carbonate, the mineral calcite, is one of the best known of the vein-filling and cementing materials. It is one of the most abundant of the constituents carried in solution in natural waters and makes up a large part of the chemical and organic sediments, the limestones, dolomites and gypsum deposits. While lime is thought of more largely in connection with the chemical and organic sediments, the above ratios show that lime is contributed in even larger amount to the fragmental sediments, the shales and sandstones. It occurs in fragmental sediments as cementing and vein-filling material, added to these rocks after their formation and as chemical or organic precipitates formed simultaneously with the deposition of the mechanical sediments.

Comparison of the amount of lime in the average igneous rock with the amount of lime in the composite of the sediments combined in the ratios above determined (see Table XIII, p. 80) shows an excess of lime in the sedimentary rocks amounting to 32 per cent of the total amount of lime in the original igneous rocks. If the lime in solution in the sea is taken into account this discrepancy is increased.

Investigation of this discrepancy leads to the suggestion that it may be due to inaccuracies in the average analyses employed. There is little reason to believe that the lime content of the average igneous rock fails to represent approximately the true average. The average lime content of the igneous rocks represents an average of 1437 determinations,[1] while the lime content of the shales, sandstones, and limestones, represents

[1] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, pp. 27–28.

only 78, 253, and 345 samples respectively. The lime content of the carbonate rocks may be said to vary from that of calcite to that of dolomite, and the average limestone analysis representing a ratio of calcite to dolomite of 56 to 36 (Table XII, p. 76) probably comes close to a true average of the carbonate rocks. It is impossible to account for the discrepancy by using a more dolomitic rock for the average limestone. (See p. 85.) The average sandstone contains only a small amount of calcium carbonate, and a maximum possible variation of lime in a reasonable average of the sandstones could not explain the large discrepancy of lime.

There seems to be more opportunity of finding the explanation of the lime excess in the shales. The average analysis [1] of the shales employed represents a composite made up of 78 samples of Paleozoic, Mesozoic, and Cenozoic shales weighted in proportion to the thickness of the formations represented. The shales make up over 80 per cent of the sedimentary rocks represented in the composite analyses, consequently a slight variation in lime content in the shale more largely affects an average of the sediments than does a variation in lime content in the limestones or sandstones. A composite of 27 Mesozoic and Cenozoic shales which enters into the average analysis of the shale has a lime content of 5.96 per cent, while the composite of 51 Paleozoic shales has 1.41 per cent lime. The average analysis of the shales fails to take into account any of the shales and slates older than the Paleozoic, which are, in general, lower in lime than the younger unmetamorphosed shales. This is apparently a reason for believing that the lime content of the average shale analysis employed may be too high and that the true average lime content of the shales may be as low as 1.41, the lime content of Clarke's composite of the Paleozoic shales. If this figure is used in computing the lime content of the average sediments in the ratio of 82, 12, and 6, the discrepancy in lime practically disappears.

Redistribution of Soda

The oxide of sodium is present in the three principal types of sediments in the following proportions: shale 95 per cent of the total soda in sediments, sandstone 4.7 per cent, limestone 0.3 per

[1] Clarke, F. W., Analyses of rocks from the laboratory of the U. S. Geological Survey: Bull. 228, U. S. Geol. Survey, 1904, p. 21.

cent. A comparison of the average analysis of the igneous rocks with the composite analysis of the sediments (Table XIII, p. 80) shows a deficiency in the sediments of 64 per cent of the soda in average igneous rocks, obviously due to the fact that sodium is concentrated in the sea and deposited with the sedimentary rocks to only a limited extent.

Soda is the most soluble of the rock constituents and it consequently is transported in solution during redistribution, except where the original rock minerals are transported mechanically in an undecomposed condition.

Under arid conditions and through dessication of ocean and other saline waters sodium salts are precipitated and sodium becomes a rock constituent. Exposure of these rocks to the surface, under moderately humid conditions, permits ready solution and transportation of the sodium to the sea. Sodium also becomes a constituent of the sedimentary rocks in so far as the sea water is included and retained by the sediments.

Redistribution of Potash

Potash in the three principal types of sediments is distributed in the following ratios: shales 93.8 per cent, limestone 5.5 per cent, sandstone .7 per cent. A comparison of the average igneous rock with the composite analysis of the sediment shows an excess in the sediments of 2 per cent of the potash of the average igneous rocks.

Potash appears to be less soluble under conditions of katamorphism than lime or soda. It occurs in relatively stable hydrous aluminum silicates of the sericite type which are transported mechanically with the clays. Furthermore, potash is characteristically retained by the clay minerals, either in combination or by adsorption, and remains to a considerable extent with these clay minerals during their transportation and deposition. To some extent potash is contributed in solution to the sea.

CHAPTER VIII

GENERAL SUMMARY OF KATAMORPHIC CHANGES AND CONDITIONS

The term *katamorphism* is used to cover all alterations of a disintegrating or decomposing nature, whether accomplished by weathering or by thermal solutions, whether at the surface or below. The most important phase of katamorphism is *weathering*, or the alteration of rocks near the surface by agencies of the atmosphere and hydrosphere.

WEATHERING

General Account

When a rock is exposed to atmospheric agencies it tends to become decomposed and disintegrated. Part of its constituents are carried away in solution; the less soluble remainder, somewhat changed in composition, forms a porous residual mass which often retains the texture of the parent rock. This is ultimately removed mechanically by the agencies of erosion. Igneous rocks, fragmental sediments containing undecomposed silicates, and the carbonate formations yield more readily to decomposition and solution than clay and quartz rocks (shales and sandstones), which are more nearly adapted to katamorphic conditions. Under certain influences the clay and quartz rocks may even undergo cementation or case hardening when exposed to atmospheric agencies, which is a process essentially constructive or anamorphic in its nature. (See pp. 121 and 129.) The end-products of katamorphism are clay, sand, carbonates, vein-filling and cementing materials, and salts of underground and surface waters. With the sands and clays are varying proportions of undecomposed silicates from igneous and crystalline rocks. The sediments are the principal end-products of katamorphism. The proportions by weight of sediments formed from average igneous and crystalline rocks, as cal-

culated in Chapter VII, Part I, are muds or shale 82 per cent, sand or sandstone 12 per cent, and lime and magnesium carbonate formations 6 per cent. The calculated proportions of sediments resulting from the katamorphism of an average igneous rock should not be regarded as necessarily the proportions of sediments actually existing to-day. (See Chapter V, Part I.)

Chemical Changes Caused by Weathering

It is apparent from the discussion of the chemical changes of common groups of rocks in foregoing chapters that katamorphism involves chiefly four chemical processes—hydration, carbonation, oxidation, and desilication. The results of hydration appear chiefly in the development of clay; carbonation is expressed in the development of carbonates; and oxidation affects chiefly the iron; desilication means sundering of silicates and separation of quartz and bases. If the mineral changes are followed through the intermediate stages, it is found that they are usually in the direction of producing these net results. There is a constant splitting off of silica, lime, iron oxide, or clay substances. In extreme cases weathering involves even the solution of free quartz and the leaching of silica from clay, leaving hydroxide of aluminum.

Water, carbon dioxide, and oxygen are the principal agents in the alteration, but sulphuric acid, hydrochloric acids and organic acids also play their part. The reactions and agents are in detail complex. In general results they may be somewhat simply summarized as above.

By taking into account the proportions of the solid mineral products formed by the weathering of an average igneous or crystalline rock, it appears that, of the three principal substances added by weathering, oxygen constitutes 12 per cent, water 24 per cent, and carbon dioxide 64 per cent. Reduced to terms of elements, oxygen is by far the most abundant element added.

Mineral Changes Caused by Weathering

Starting with average original igneous and crystalline rocks, which are composed of about 50 per cent feldspar, 25 per cent ferromagnesian minerals, and 20 per cent quartz, the processes of katamorphism ultimately produce a combination of sediments in which quartz, carbonates and clay are the characteristic prod-

ucts, mingled to varying degrees with intermediate products of alteration or parts of the original rock mass which have been carried through the alteration without much change. The average mineralogical composition of the sediments, as calculated from average analyses, indicates that in order of abundance the minerals are quartz, white mica, feldspars, carbonates, kaolin, chlorite, and iron hydrates.

There has been a simplification of minerals. The original silicates, which made up 75 per cent of the igneous and crystalline rocks, constitute only 16 per cent of the average sediments and in extreme cases of weathering entirely disappear. The destruction of feldspar is the most important single mineral change. The bases of these silicates have been segregated largely in carbonate and iron oxide deposits. Soda has gone largely into the ocean.

During the process which tends to develop the end-products there are a considerable number of silicates formed which are to be regarded as intermediate stages in the alteration. For instance, feldspars seldom go directly to clay with a complete separation of the bases, but intermediate products, like sericite and epidote, are common developments. When any of these minerals develop from the alteration of a feldspar they do so with a separation of some quartz or calcite or kaolin (p. 4). Likewise the original pyroxenes seldom go over directly to the end-products. They are likely to show a change first to hornblende and chlorite. Hornblende in turn breaks down into chlorite. The alteration of pyroxene to hornblende involves the separation of calcite, and that of hornblende to chlorite likewise involves this separation. In studying rock alterations under the microscope or from calculation of analyses, the results can be best correlated and accounted for if it is remembered that the tendency for alteration is always towards certain simple end-products. This view of the alteration reduces to order a vast number of complex alterations which would otherwise have no general significance to the student.

Volume Changes Caused by Weathering

In the discussion of the weathering of acid and basic igneous rocks on previous pages, it appears that katamorphism increases the volume: (a) by mechanically disintegrating and separating

materials, forming openings; (b) by the addition of water, carbon dioxide and oxygen, and (c) by decreasing the average specific gravity of the minerals. This does not mean that minerals of higher specific gravity may not be formed, as for instance limonite, but the average density is lowered when all the resulting minerals are taken into account.

It is of interest to note that the principal substances added,—oxygen, hydrogen, and carbon,—are all of light weight. It is to be expected, therefore, that the molecular combinations to which they enter should be on the whole lightened and made less dense.

By comparing the relative weights and densities of the average original rock and the resulting sediments, it has been possible to get some idea of the magnitude of the volume change involved. The net increase in volume in the change from average igneous rocks to average unconsolidated sediments amounts to 54 per cent of the volume of the original rock. (See Table IX, p. 71.)

Conditions Favoring Katamorphism by Weathering

In the foregoing pages we have discussed chemical, physical and mineralogical changes in weathering as actually observed or measured in the rocks themselves. Reference has also been made to the chemical agencies in these changes. It remains to discuss the conditions and environment which have determined the observed alterations.

Igneous rocks are most readily altered at or near the earth's surface. Sedimentary rocks are themselves derived ultimately from igneous rocks by katamorphism, and may themselves be regarded essentially as the products of katamorphism, on which further katamorphic action proceeds far less effectively than on the igneous rocks.

When an igneous rock is brought to the surface of the earth it is subjected to a new set of physical conditions. The agencies of erosion cause mechanical breaking and disintegration. These physical changes are one phase of the adaptation of the rock to its new environment.

The rock also becomes surrounded by a new set of chemical conditions. Its minerals come in contact for the first time with the substances of the hydrosphere and atmosphere, and some of their constituents combine with carbon dioxide, water, oxygen,

or other substances, to form new minerals better adapted to the new conditions.

The first and must important condition of katamorphism, then, is the accessibility of these substances to the rocks. In general this accessibility is at a maximum near the surface and decreases in going downward. The zone through which katamorphism extends is called the *zone of katamorphism.* This, however, has very indefinite limits. At very moderate depths both katamorphic and anamorphic changes may occur simultaneously in different rocks or in the same kind of rocks under different conditions.

Above the water table, in what is usually called the *belt of weathering*, the conditions are more favorable for katamorphism than below the water table. Above the water table there is opportunity for the interaction of all three of the principal agents, oxygen, carbon dixoide and water, in both gaseous and liquid form, whereas below the oxygen and carbon dixoide are much less abundant. Above the water table temperature changes are greater, thereby aiding both the mechanical and chemical changes of the rock. All the agencies of erosion tend to disintegrate the rocks, separate their parts and give access to the chemical agents. There is here the room necessary for expansion of volume during weathering. Plants and animals play an important part in the chemical and mechanical changes. The chemical changes brought about by plants are in detail complex, but their general result is clearly carbonation.

The thickness of the belt of weathering ranges from zero at water level to a thousand feet or more in certain arid regions of high relief. In speaking of the depth of the belt of weathering we refer to the places where weathering processes are at work, and not necessarily to depth of the residual mantle of products of rock weathering, for within the belt where weathering processes are at work the residual products of rock decay may be many or few, depending on a variety of local conditions. For instance, in certain arid countries of high relief, where the water table is far below the surface, the distinctive products of weathering may appear only along fissures, the products formed at the surface being removed by the rapid erosion about as fast as they are formed. Likewise in areas of recent glaciation igneous and crystalline rocks

may show only the beginnings of weathering, notwithstanding the fact that the process is operating along fissures at least down to the water table. In some cases less than an inch of the rock near the surface or near fissures has been conspicuously affected since glacial scouring. On the other hand, in flat-lying regions of warm humid climates and heavy vegetation the rocks may be practically entirely decomposed down to, and even below, the water table. A depth of 200 feet of complete decomposition is not unusual in tropical regions. Development of a thick mantle of residual weathered material requires not only favorable conditions for decomposition, but an absence of conditions favorable for the rapid erosion of these products. Even in a tropical country, in which decomposition is rapid, there is only a small accumulation of weathered materials along steep slopes where erosion is active.

It is generally assumed that decomposition is less active in arid climates of high relief, for the reason that there is usually no thick mantle of rock decay and such mantle as remains shows little decomposed silicates. This does not prove that the speed of decomposition is any less under such conditions than in tropical regions of high humidity, for erosion follows decomposition so closely that there is not time for the accumulation of decomposition products even though the total amount of chemical change on a given unit of rock may be as rapid as in a humid country. From *a priori* reasoning there is ground for believing that decomposition in an arid country is not so rapid as in a humid country, because of the absence of vegetation, implying absence of carbon dioxide and lack of uniform supply of water, which are such important agencies in the alteration. The same conclusion is indicated by the estimate that some rivers in arid countries carry less dissolved mineral matter per square mile of drainage area per year than rivers of humid climates. (See p. 254.)

Not only does the degree of weathering vary with climatic and topographic conditions, but many other factors influence the response of rocks to weathering. Basic igneous rocks weather more readily than acid igneous rocks, due to the fact that the agents of weathering are largely acid. Coarse-textured plutonic rocks weather less rapidly than surface flows. Either weathers less rapidly than fragmental tuff or ash. Sandstones and shales are

likely to show less weathering than any of the igneous rocks. Limestone may be more readily taken into solution. Secondary deformation of rocks, causing openings, may locally promote weathering. The large quantities of sulphuric acid freed by the oxidation of sulphide deposits locally alter rocks to a large extent.

Below the water table the processes of katamorphism are less conspicuous than above, but they still cause changes the same in kind. Less material goes into solution and more is deposited in this zone than in the upper belt of weathering. Locally katamorphism may be extensive for hundreds of feet below the water table. The essential condition for this is the free circulation, through continuous openings, of waters carrying the agents of alteration, opportunity for the necessary increase in volume, and that the agents of alteration shall not be abstracted from the solutions by alterations near the surface. Oxygen may be abstracted near or above the water table, but where carried down in contact with rocks which are not abstracting oxygen, it may go a considerable distance below the water table. As reducing agents higher up become oxidized, the oxygen is able to travel farther down. Thus, oxidizing solutions have travelled many hundreds of feet below the present water table in the Lake Superior iron formations, and probably far below any water table of the past, for secondary oxides are found at a depth of at least 2,000 feet. Presumably in the early stages the oxygen was largely abstracted from the solutions by the oxidation of iron carbonates which the waters met. But as the zone of iron carbonate near the surface became oxidized the oxygen-bearing waters were able to travel through it and accomplish their work farther down. Carbonic and sulphuric acids may accomplish their work near the top of the water table, or they may be carried to considerable depths before being neutralized. In proportion as the rocks are fragmental end-products of katamorphism, they allow the katamorphosing solutions to pass through them with less change to greater depths.

At any depth at which katamorphic processes are active, anamorphic processes may be simultaneously working—even in the belt of weathering. Anamorphic processes of cementation and recrystallization may be integrating certain rocks like the fragmental sediments while the igneous rocks are being decomposed. The same is true to a more important extent below the belt of

weathering. Certain soft rocks may be locally yielding to the process of rock flowage, which is essentially an anamorphic process, at the same time that adjacent rocks are still being katamorphosed. In the vicinity of igneous intrusives anamorphic processes may be effective, whereas a short distance away the process may be dominantly katamorphic.

Further observations on the conditions of weathering are made in the discussion of river waters in Chapter II, Part III.

The solutions effective in katamorphism are mainly of surface meteoric origin, of ordinary temperature, carrying carbonic acid or locally sulphuric acid. This is shown by actual tests of waters observed to cause the alteration, by the mineral products developed, and by the localization of katamorphism to points accessible to such waters. The altering waters are essentially downward and laterally moving. Meteoric waters may move upward through the zone of katamorphism, but they are less effective in altering rocks in this part of their course for the reason that they have largely lost the active acid agents of alteration in their downward and lateral course. Locally katamorphism may be caused by hot solutions from igneous sources.

KATAMORPHIC ALTERATIONS OF IGNEOUS ROCKS BY THERMAL SOLUTIONS

Under moderate temperatures and pressures igneous rocks may be altered by thermal solutions in an essentially katamorphic manner. The locus of such alteration may extend from the surface to a depth of perhaps 8,000 feet, as shown by observation of the alterations in relation to erosion. The essential condition is accessibility to the rocks by the hot solutions. Thus it is that these alterations are confined to fissures and other openings below the surface in the vicinity of igneous rocks, or to the vicinity of volcanoes at the surface. Near volcanoes the waters may be downward, lateral, or upward moving when they affect the alteration. Below the surface the waters have ordinarily been regarded as essentially upward or lateral moving, deriving their heat from igneous intrusives or extrusives, probably also deriving some of their contained salts from the same source, and perhaps being themselves at least in part of magmatic origin. At the surface, in the vicinity of volcanoes, the waters are usually distinctly acid,

possibly due to contact with the atmosphere. Below the surface the character of the solutions may be inferred only from the alterations accomplished, or locally from the content of hot springs. These waters are usually of alkaline rather than of acid reaction, and are often more concentrated than surface solutions. The spring waters in the vicinity of recent volcanic activity contain hydrogen sulphide, or sodium chloride, locally containing abundant silica, or sodium carbonate generally associated with free carbon dioxide, and may locally contain notable amounts of sodium sulphates. Comparison of analyses of the fresh and altered products of thermal alteration indicates the highly varying character of the waters. Usually there is evidence of introduction of sulphur, carbon dioxide, potash, silica, and locally metallic ores.

The minerals developed by hot waters include all those characteristic of weathering, and certain other minerals in addition. Most characteristic of hot waters are sericite, adularia, quartz, chlorite, epidote, and carbonates.

ENERGY CHANGES IN KATAMORPHISM

The principal chemical changes which accompany the katamorphism or decomposition of an igneous rock are hydration, carbonation, and oxidation. In regions of abundant sulphides sulphuric acid may be present in the solutions, and sulphidation rather than carbonation may be important. All of these reactions are characteristically exothermic, that is, energy in the form of heat is liberated as the reaction goes on. When the heats of formation of the several compounds involved in a chemical reaction are known, it is possible to write the heat equation and determine the energy changes in calories (pp. 305–306). The heats of formation of a considerable number of mineral molecules have been determined experimentally with a fair degree of accuracy, but there are such gaps in the quantitative information of this kind that it is not yet possible to make a really adequate statement of the heat developed in the katamorphic alterations of a fairly simple rock like a granite, although this may be approximated by computing the heat developed in the oxidation of a ferrous iron, the carbonation of lime and magnesia, the hydration of alumina, and the chloridation of soda and potash.

In general, solution of substances absorbs energy and precipi-

tation liberates energy. Solution and precipitation may be supposed to approximately balance during katamorphism, except for the substances remaining in solution in the ocean. So far as this factor is concerned, the net result may be a slight absorption of energy, but this is small as compared with the liberation of energy from chemical reactions.

The net result of katamorphism seems clearly to be liberation of energy, though part of it is reabsorbed by solution.

PART II

ANAMORPHISM

ANAMORPHISM DEFINED

Under anamorphism are considered all of the integrating and constructive changes in rocks, tending to make them coherent and crystalline. The dominant mineralogical feature is the secondary growth of silicates. The term *anamorphism* will be used to cover: (1) cementation, including all induration and consolidation by infiltration of cements, by recrystallization, or by mechanical compression; (2) dynamic metamorphism or rock flowage; and (3) contact and thermal metamorphism. It includes also in each of these classes certain "metasomatic" replacements by substances introduced from without, so far as these tend in the general direction of anamorphism. Metasomatism is often treated as a separate division of metamorphism, but metasomatic processes are so closely involved with other types of metamorphic changes that we find it more convenient to treat it as a part of these changes.

CHAPTER I

ANAMORPHISM OF CLAYS (ARGILLACEOUS ROCKS OR PELITES)

CEMENTATION OF CLAYS

Nature and Source of Clays

The term clay is here used in a general sense to cover the argillaceous products of weathering.

The primary source of clays is the igneous rock. The argillaceous sediments constitute 82 per cent of the end-products of katamorphism of igneous rocks. Small percentages of clay constituents are deposited with the other sediments. Clay is produced by decomposition of sedimentary rocks, especially limestone and shale. Consolidated clay and sand formations are reasonably resistant to decomposition, but yield a considerable amount of clay where chemical processes are aided by disintegration.

Clays represent the fine-grained products of rock decay and disintegration, which have been separated from the coarser-grained materials by the sorting which accompanies erosion, transportation, and deposition. They also include certain substances deposited from solution. The fine-grained products of rock decomposition are the clay minerals of the kaolin group, hydrous iron oxides, such as limonite, and fine-grained hydrous aluminum silicates of ferrous iron and magnesia of the chlorite type. Sericite is known to develop from the feldspars during katamorphism, and so far as this mineral escapes further alteration to kaolin, it may be contributed to the clays and muds. In the clays also an impure white mica, sericite, or muscovite, may develop secondarily. Muscovitic decomposition, according to Hutchings[1] is even more usual than kaolinitic decomposition.

[1] Hutchings, W. M., Notes on the composition of clays, slates, etc., and on some points in their contact metamorphism: Geol. Magazine, vol. 1, 1894, p. 40.

The fine-grained products of mechanical disintegration of rocks may be deposited with the clays as finely divided quartz, feldspar, ferromagnesian minerals, and mica. Because of their fine state of division they are probably easily altered during water transportation, and even subsequent to deposition. Of the soluble products of rock decay, the carbonates of lime and magnesia are deposited with the clays to a limited extent, largely by organisms. Soda and potash carried in solution may be absorbed by the clay minerals, particularly those in the colloidal form. Colloidal silicic acid is believed to be deposited largely with the clays and muds, as the same processes of flocculation which precipitate the colloids also throw down the finely divided sediments. Sulphates are deposited with the clays in small amounts. The minor accessory constituents of rocks, such as magnetite, hematite, garnet, zircon, ilmenite, apatite, etc., are contributed in small amounts.

To summarize,—the minerals of the clays consist of hydrous aluminum silicates of the kaolin group, containing more or less of adsorbed alkalies and alkaline earths, minerals of the chlorite group, hydrous iron oxide, secondary quartz, and opal, carbonates of calcium, magnesium, and possibly iron, small amounts of sulphate (probably largely gypsum), fine-grained fragmental quartz, feldspars, micas, and ferromagnesian minerals, and very small amounts of original unaltered accessory minerals. Small amounts of carbon, phosphoric acid, and titanic acid in the clay analyses evidence the presence of carbonaceous material, phosphates, and titanium minerals. It is probable that the phosphoric acid in clays is present in secondary calcium phosphate rather than original apatite of the igneous rocks.

Shales Resulting From Consolidation and Cementation of Clays

The consolidation and cementation of clay produces shale, which may have an excellent parting parallel to the bedding, or may be nearly massive. It is difficult to follow specific clays or shales through this change, but a comparison of the clays as a group, with shales as a group, shows what has been accomplished during this process. The processes of induration of clays will be considered under the general term cementation.

Locus of Cementation of Clays

The extent to which the process of cementation may go in the early life of a clay is as yet largely problematical. As is the case with other sediments, it has usually been assumed that cementation occurs principally after the sediment becomes buried and lies for a long time beneath groundwater level. The possibility of early cementation, beginning at the moment the clay is deposited, has not received sufficient consideration. Soon after deposition, as in growing deltas, layers of clay acquire sufficient competency to transmit stresses, and to develop folds and faults. Mud lumps and clay masses of considerable solidity are observed to form under water on muddy shores and deltas. Drying and settling of clays lead to consolidation from the start. We may conclude that the consolidation of clays ordinarily begins at the moment of deposition, and continues long after burial.

Conditions and Agents of Cementation of Clays

Argillaceous sediments, although very porous, are characteristically impervious and do not permit even moderately free circulation of solutions. Water circulation depends on size and continuity of openings rather than on total volume of openings or degree of porosity. Cementation by infiltration of materials from extraneous sources is believed to be largely inhibited by the imperviousness of clay to free water circulation. Induration is probably accomplished mainly by compression and internal rearrangement of original constituents. Elimination of soluble constituents is even probable when water is squeezed out by compression. If there has been infiltration of cements, this is difficult to prove microscopically because of the extremely fine texture, nor can it be proved by comparison of analyses of clays and shales. The major causes of cementation are believed to be of another sort, described below.

Clay is very rapidly hardened and cemented by air drying. Pressure accomplishes similar results, as shown in brick making. This hardening is due to elimination of free water and mechanical packing, accompanied by dehydration of gelatinous aluminum silicates, organic colloids, gelatinous silicic acid, ferric hydrate,

and in some cases aluminum hydrate.[1] The cohesive property of clay has been attributed to colloidal envelopes surrounding minerals. An excess of non-colloidal minerals, such as quartz and mica, reduces the cohesiveness. Flint, hornstone, and amorphous chert, often found in shales, are regarded by some investigators as forms of silica necessarily deposited from a solution of a colloidal nature, possibly associated with organic agencies. Consideration of the distribution of the end-products of weathering suggests rather strongly that colloidal silica is deposited with the transported clays and in fact may be a cause of settling of the clays, as white of egg settles coffee.[2] These colloidal substances probably have much to do with the plasticity of clays, although this has also been ascribed to the combined water, and to the shape and size of the constituent particles. According to Ashley,[3] the adsorption of a dye by the colloids of a clay supplies an approximate measure of plasticity. In the early stages of cementation, before much crystallization and dehydration have taken place, the process of hardening may be reversed by soaking and disintegration in water. This, however, may not completely restore the original degree of plasticity. Apparently from the beginning of consolidation some crystallization occurs which is not easily reversible.

Minerals of Shales

The average mineral composition of shales is given on page 76. There are of course wide variations from this average. The shales have a much smaller proportion of amorphous constituents than the clays. The colloidal substances have largely disappeared. Kaolin and the ferric hydrates have decreased to a marked extent. The fine-grained matrix or paste of the shales consists of finely granular quartz or chert, containing considerable quantities of white mica and locally rutile. Much of the iron is reduced to the ferrous condition in combination with silicates (principally chlorite) and as carbonate. It may be combined with sulphur,

[1] Ashley, H. E., The colloid matter of clay and its measurement: Bull. 388, U. S. Geol. Survey, 1909, pp. 9–13.

[2] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey. 1911, p. 482.

[3] Op. cit., p. 59.

which is often present under these conditions, to form pyrite. The presence of carbonaceous matter favors this reduction, as shown by the common association of iron carbonate and sulphides with graphitic shales. The result is often a change of color of the rock from yellowish or reddish to greenish gray or black. Where organic material is absent, as in certain terrestrial muds, lack of deoxidation may permit the characteristic reddish and yellowish colors to remain.

In describing the constituent minerals of the argillaceous rocks, Harker [1] makes the following statement:

"The ordinary fine-grained argillaceous rocks consist in considerable part of an exceedingly fine-textured base or paste, very difficult to resolve, in which any truly detrital elements or their evident alteration-products are imbedded. The mature of this paste has not yet been made out in any large number of cases. It was formerly regarded as consisting essentially of hydrated silicate of alumina (kaolin, etc.). Careful studies of various clays, shales, and slates lead, however, to the conclusion that the material is to a great extent finely divided white *mica*, and it is now generally admitted to be of secondary origin. With this is much indeterminable, finely granular matter, which may be conjectured to represent the finest powder of quartz, feldspar, etc., and perhaps *kaolin* or other products."

Hutchings [2] has investigated the clays, shales, and slates in considerable detail. He finds that white mica develops abundantly in very early stages of cementation when the rock is still soft enough to be easily cut with a knife. It is very fine-grained and intimately intermixed with the other constituents. Also, it contains, in addition to potash, much magnesia, iron, and soda, as shown not only by analyses but also by heating the material and thus making the iron conspicuous by oxidation. He reports further that with advancing metamorphism the muscovite becomes more coarsely crystallized and tends to eliminate its impurities, principally magnesia and iron, with the result that chlorite begins to develop. When the slate or schist phase of metamorphism is reached, the mica is definitely muscovitic and well separated from the chlorite, which becomes more or less conspicuous

[1] Harker, A., Petrology for students, 1895, pp. 208–209.

[2] Hutchings, W. M., Clays, shales and slates: Geol. Magazine, vol. 3, 1896, pp. 309 et seq.

in this stage of metamorphism.[1] The chlorite in the clays is mainly clastic. During the early stages the muscovite tends to develop more or less parallel to the bedding, but later, under dynamic and contact metamorphism, it may cross the bedding in definite planes, forming cleavage, or at igneous contacts may form a felted mass.

The abundant development of muscovite in the matrix or paste of a shale is proved not only by observation but by chemical analysis. In calculating the mineral composition of shales from chemical analysis, a lack of sufficient water to combine with alumina as kaolin makes it necessary to combine part of the alumina, alkalies, and water in mica molecules. The probability of considerable mica in the shales is also evidenced by the ease with which visible mica flakes develop in argillaceous rocks under dynamic metamorphism. As mica is present in extremely minute particles in the shales, recrystallization alone without rearrangement of constituents during dynamic metamorphism is sufficient to develop the mica in slates and phyllites.

The existence of so much mica in shales presumably finds its explanation in the remarkable affinity for potash of colloidal clays and silicates, which has been determined experimentally and which is clearly shown by the retention of potash in residual clays resulting from weathering. These potash-rich collodial clay minerals resulting from weathering have practically the same silica-alumina ratio as muscovite, with a higher content of water. The large development of muscovite in comparatively early stages of cementation suggests strongly that this is merely a case of partial dehydration and recrystallization, perhaps due largely to pressure. In the discussion of weathering it has been noted that muscovite or sericite may develop directly from the weathering of feldspar or other silicates, without going through the kaolin stage. There are several ways of considering this—the muscovite may be regarded as an intermediate product of weathering between the unaltered silicate and the clay, or it may be the end-product which is prevented from going to clay by pressure, or it may be the end-product which develops when there is a dearth of carbon dioxide or other acids to remove bases. There is evidently a very delicate balance between muscovite and colloidal clay containing potash

[1] Op. cit., p. 312.

during the later stages of katamorphism and the earlier stages of cementation.

The local abundance of rutile needles in the fine-grained matrix or paste of a shale, sometimes called "clay slate needles," presents a problem similar in many respects to that of the genesis of muscovite. Some rutile is present in clays as fragmental particles from primary rocks. Some rutile has been observed to develop secondarily by katamorphism, especially from biotite, sometimes from amphibole and olivine. The largely increased quantity of rutile in the shales indicates recrystallization of titanium compounds during cementation. The particular source of the titanium for this recrystallization is not entirely clear. It has been ascribed to the alteration of leucoxene, an amorphous mineral, formed by weathering of ilmenite. The composition of leucoxene is doubtful but it has been described as approximating titanite. The close associations of the rutile with the muscovitic and siliceous matrix or paste of shales, and the fact that it is contained in this paste suggest the possibility that the colloids in the clays, which have recrystallized to form this paste, contained titanic acid, adsorbed in the colloids, just as potash is adsorbed. There have been comparatively few tests for titanic acid in the colloids, and such adsorption has not been experimentally determined. Clarke[1] publishes an analysis of "a clay with the silica replaced by titanic oxide" from a weathered rock in North Carolina. To this mineral the name *xanthitane* has been given.

It is clear that temperature changes do not have much to do with the development of rutile and muscovite, for these seem to occur essentially as processes of consolidation of clays near the surface. It seems reasonable to suppose that slight variation in pressure may be the controlling factor.

There remains to be mentioned another method of cementation which may be operative in the shale, namely, the enlargement of original fragmental crystals of quartz and some silicates in the manner described in connection with the cementation of sands. (See pp. 123–124.) Presumably there is enlargement in the shales as in the impure sandstones and gradational phases between sand-

[1] Clarke, F. W., Report of work done in the division of chemistry and physics mainly during the fiscal year 1887–8: Bull. 60, U. S. Geol. Survey, 1890, p. 135.

stone and shales, but the texture is so fine that it is difficult to observe.

The essential mineralogic changes in forming shale from clay are graphically summarized in Fig. 11.

Volume Changes in Cementation of Clays

The settling of the clays, and particularly drying, involves a reduction of volume. The mineral and chemical changes are too imperfectly known in this stage of the process to answer definitely the question as to what extent they may change the volume. The change from alkali-bearing kaolin to white mica certainly involves reduction in volume, due both to loss of water and increase in specific gravity. It has been calculated (see Table IX, p. 71) that the average shale has a mineral specific gravity of 2.71 and that a corresponding clay has a mineral specific gravity of 2.68, which, together with dehydration, accounts for a decrease in volume of 1.5 per cent, neglecting pore space. Porosity is largely eliminated by compression. The average porosity of clays is 27 per cent while that of shales is 13 per cent. Including decrease in porosity, the decrease in volume from clay to shale is 17 per cent.

ANAMORPHISM OF CLAYS AND SHALES INTO SLATES

A clay or shale becomes a slate through dynamic action or contact metamorphism, possibly even by the effects of high temperature and pressure, independent of either, or possibly by high pressure alone. The term slate is ordinarily applied to an argillaceous rock having secondary cleavage, but it is difficult in practice to restrict the term closely to this type of rock. Argillaceous rocks, under the influence of intrusion, may be converted directly into contact rocks devoid of cleavage which are mineralogically and chemically slates. Extreme induration and recrystallization of a shale, independent of dynamic action or contact metamorphism, may produce a highly crystalline rock without cleavage, to which the term slate is usually applied. In degree of metamorphism and composition these rocks are essentially similar to the slates and not to the shales. It is not convenient or necessary to apply any other term than slate to these rocks.

Anamorphism of Clays and Shales Independent of Dynamic or Contact Action

Deep burial of a clay or shale formation may develop a highly crystalline rock, mineralogically and chemically allied to the slates. The changes are essentially a continuation and completion of those involved in the change of clay to shale,—principally further dehydration and an increase in amount and coarseness of crystallization of the mica. Whether these conditions can simulate the results of the more intense anamorphism of dynamic and contact action is open to question. Where the evidences of contact action are marked and also in slates which do not give evidence of contact action, crushing or shearing, the effects of a long continued submission to high temperature and pressure, due to deep burial, cannot be excluded. Quoting from Hutchings,[1]

"we cannot here deny the possibility that what we see is the result of simple depth conditions. . . . We can only leave this question an open one, awaiting further evidence; and simply note the fact, that where we have these cases of high development, not directly accounted for, they are always on the same lines exactly as in contact areas, and are often plainly not dynamic."

Dynamic Metamorphism of Clays and Shales Producing Slates, Schists and Gneisses (Rock Flowage)

MINERAL, CHEMICAL, AND TEXTURAL CHANGES IN FORMING SLATE FROM SHALE BY DYNAMIC METAMORPHISM. The mineralogical, chemical and textural changes observed in the production of the shales from clays, continue under dynamic metamorphism. (See Fig. 11.) Mineralogically and chemically the slates resemble the shales. The specific effect of dynamic metamorphism is the development of a parallel arrangement of mineral constituents, which imparts a fine cleavage to the rock, generally at an angle to the bedding. As cleavage is common to other rocks, a separate chapter is devoted to its discussion. (Chapter VI, Part II.) The bedding may still appear as color bands or bands of differing composition which are frequently rendered conspicuous by weathering. The change is marked by the more conspicuous development of platy minerals, principally muscovite, and to a less extent chlorite,

[1] Hutchings, W. M., Clays, shales and slates: Geol. Magazine, vol. 3, 1896, p. 347.

which are mainly responsible for the cleavage of the rock. These minerals may constitute as high as 30 to 40 per cent of the slates and only a few per cent of the shales. An increase in quantity of the mica marks a further stage of this process, when the phyllite is produced, in which mica may constitute more than 50 per cent of the weight of the rock.

According to Hutchings,[1] the recrystallization of the muscovite causes elimination of impurities, resulting in the conspicuous appearance of chlorite in the later stages of metamorphism. In this stage, also, biotite sometimes begins to develop in the interstices of the muscovites, indicating its contemporaneous crystallization. Hornblende also locally develops with parallel arrangement, but this mineral on the whole seems to be more characteristic of the contact metamorphism of slates than of dynamic metamorphism.

Groups of analyses specifically representing the change from shale to slate are lacking, but a comparison of the analyses of the shales as a group with the slates and schists as a group, shows insignificant changes. Important change in composition is not to be expected, for the reason that the shales and clays approach the composition of the average slates. There are present all ingredients which are necessary for the production of the silicates characteristic of the slates, and the materials in excess of this requirement are not relatively abundant. The change from clay to shale produces a rock which is mineralogically and chemically well toward a slate in composition. While some further mineral and chemical changes occur in development of slate, these are relatively slight and subordinate to the change in texture.

MINERAL, CHEMICAL, AND TEXTURAL CHANGES IN FORMING SCHIST AND GNEISS FROM SHALE AND SLATE BY DYNAMIC METAMORPHISM. Schists and gneisses are known often to develop under extreme conditions of anamorphism, which suggests that they may represent stages of anamorphism beyond that of the shale, although it does not follow that all schists and gneisses which have been produced from shales have gone through the intermediate stages of slates or phyllites.

In general the schists represent merely a continuation of the

[1] Hutchings, W. M., Clays, shales, and slates: Geol. Magazine, vol. 3, 1896, pp. 312-313.

recrystallization processes which developed the slate, resulting in coarser grain and consequent increase in visible mica.

The gneisses are characterized by the presence of feldspar and in general have a coarser texture than the schists. They are believed to represent a still more advanced stage of anamorphism in which the micas and chlorite have in part been altered by dehydration to feldspars and hornblende. These changes are frequently accompanied by development of anhydrous aluminum silicates and corundum, which represent the constituents of the micas not required by the feldspars. The essential chemical change is loss of combined water.

According to Grubenmann (see pp. 188–189) a clay sediment is characteristic of an upper zone of metamorphism, a mica schist of a middle zone, and a gneiss of a lower zone. The writers have been unable to check this with field observation. Depth is only one of the factors determining intensity of conditions. Igneous intrusion, mineral and chemical composition, the differential stress conditions, etc., all play their parts. It probably would be unsafe to assume that phyllites, mica schists, and mica gneiss always develop in this order with increasing intensity of anamorphic conditions, because of the variety of factors which determine the nature of the anamorphism.

SECONDARY PORPHYRITIC CONSTITUENTS FORMED UNDER DYNAMIC METAMORPHISM. Phenocrysts of anhydrous silicates such as garnet, staurolite, andalusite, chloritoid, cordierite, albite, rutile, tourmaline, ilmenite, etc., begin to develop in the slates and phyllites and may become conspicuous in the schists. They are usually, though not always, developed after the cleavage has been essentially formed and after movement of the rock mass has ceased, because their arrangement is in diverse directions, crossing the cleavage at all angles, they do not disturb or distort the rock cleavage, and they have usually not suffered the crushing which would result from movement after their formation. The phenocrysts appear to have used the substances available at the place of development. The substances not needed may remain as inclusions, which not infrequently lie with their longer axes parallel to the cleavage. The fact that these phenocrysts are uniformly anhydrous minerals with a higher density than the average of the adjacent rock, and the fact that they are developed in the

extreme phases of anamorphism of the clays, point to the inference that they were developed as result of the requirement for diminution of volume in the rock mass. (See p. 180.)

Similar developments are characteristic of igneous contacts and it is frequently difficult to determine whether they are the result of dynamic or contact action, or both.

FIG. 10. Photomicrograph of slate from Black Hills showing porphyritic development of chloritoid after movement has stopped.

VOLUME CHANGE IN DYNAMIC METAMORPHISM OF SHALE. In the development of a slate from a shale the volume is reduced by the practically complete elimination of pore space. Experimental determinations show that ordinary commercial roofing slates have, as a rule, but a fraction of 1 per cent of porosity. Specific gravity ranges from 2.7 to 2.85. The specific gravity of the shales, as calculated from the minerals, is about 2.7. The difference therefore in mineral density is slight. Comparison of the analyses of slates and shales shows also that in the change from one to the other there has been slight, if any, elimination of water, car-

bon dioxide, and oxygen; hence there is practically no diminution in volume from this cause. The total change in volume, therefore, is essentially measured by the diminution of porosity, and is approximately 12 per cent on an average.

LOCUS OF DYNAMIC METAMORPHISM OF SHALE AND SLATE. The controlling factor in the dynamic anamorphism of shales is movement. No matter how great the pressure, if there is not movement it is doubtful whether a slaty or schistose structure is developed. Wherever the differential stresses surpass the resistance of the shale, rock flowage ensues, with a resulting development of slaty cleavage. Such condition is not necessarily dependent on depth. Shale, being one of the weakest of the formations, is likely to yield more readily than adjacent formations; in fact, the differential movement of adjacent formations is largely made possible by the easy yielding of the shales. It is not at all unusual to find slates much sheared by dynamic action, while adjacent rock, such as quartzite, limestone, or igneous rocks, are massive or only fractured. To the depth to which our observation extends, either by drilling or erosion, the conditions are those of combined fracture and flow, as indicated by the close association of schistose and massive beds. For this reason dynamic anamorphism of slates may occur under the same conditions which may be katamorphosing an adjacent rock like a granite.

It has ordinarily been assumed that at great enough depth all rocks must flow, but this involves the assumption that gravitational pressure there results in movement, and it involves further the assumption that cubic compression does not so greatly add to the strength of a rock as to make it possible for it to resist any differential stresses which might be there developed. The stresses producing rock flowage are ordinarily assumed to be largely tangential, of the kind producing orogenic movements. Whether vertical gravitational pressure alone directly produces movement is difficult to prove. The lateral support would tend to prevent movement, but locally there may well be a lateral spreading as the result of gravitational stresses. In other words, shale may conceivably be buried to great depth, without undergoing movement necessary to produce a good slaty cleavage. The existence of shale beds lacking cleavage in sediments which have been deeply buried may be thus explained.

In the vicinity of intrusions, particularly those of a plutonic sort, differential stresses are set up which may result in rock flowage and the development of slates. In these cases it is difficult to discriminate between the movement locally developed by the intrusion and general orogenic movements accompanying the intrusions.

Thermal and Contact Metamorphism of Shales and Slates

The changes thus far described are essentially those of dynamic anamorphism. They are the kinds of changes that may be seen in great slate formations which have been folded and buried, without evidences of contact metamorphism. Ordinarily, however, it is practically impossible to distinguish the evidences of extreme anamorphism by dynamic agencies from those caused by igneous contacts. The descriptions given above for the later stages of anamorphism will apply almost without change to igneous contacts, but in addition there are certain features of anamorphism which are peculiar to the igneous contacts. On the whole, anamorphism seems to go farther at these igneous contacts, developing coarser grain and more conspicuous porphyritic constituents.

THERMAL METAMORPHISM OF SHALES AND SLATES. The alteration of slates by hot waters along veins and at the surface follows the same general lines as the alterations of igneous rocks under similar conditions, especially where the slates contain a considerable quantity of undecomposed silicates from the primary rocks. Near the surface in the vicinity of volcanoes the hot solutions may soften and kaolinize the slates, segregate the iron and clay, and may introduce sulphates. These changes are essentially katamorphic, but to some extent they may be of anamorphic nature. Beneath the surface they are essentially anamorphic. To moderate depths beneath the surface there may be silicification, sericitization, and carbonization along veins. Along veins at higher temperatures and pressures (or greater depth) hot solutions may in addition develop tourmaline and albite, with the destruction of the micas.

CONTACT METAMORPHISM OF SHALES AND SLATES. Near dikes and surface flows the slate may show only slight contact metamorphism, being baked into a hornstone or silicified. Near con-

tacts of large plutonic igneous masses, the slate may be more extensively metamorphosed.

Mineral and textural changes. It is not always possible to distinguish the effect of the more intense contact metamorphism from that of dynamic metamorphism. The two may be closely related. Under contact metamorphism the rocks may become slaty or schistose and all the processes described under dynamic metamorphism more or less accentuated, or the cleavage may be destroyed, and the micas develop without parallel arrangement. As a whole, the grain becomes coarser, and, so far as a parallel structure is developed, it is of a schistose or gneissic rather than a slaty type. Hornblende develops rather more freely than in dynamic metamorphism. Characteristic of the contact effect of igneous intrusion on the slate is the development of certain spotted, porphyritic, and banded textures. Albite, cordierite, staurolite, biotite, and andalusite, sillimanite, garnet, tourmaline, chiastolite, rutile, and other anhydrous aluminous silicates of this type, are characteristic developments as porphyritic constituents. Graphite, epidote, chlorite, amphibole, and other minerals may be developed in groups, or spots, or bands, yielding rocks known as spilosite, desmosite, knotenschiefer, etc. These processes may be accompanied by the formation of a dense hornstone matrix.

Carbonaceous matter in the shales is one of the first ingredients to be changed by contact metamorphism. It is either dissipated and expelled or converted into graphite, which may be aggregated into little dark spots producing spotted slate or knotenschiefer. This disappears with advancing metamorphism. According to Harker:[1]

"The minute needles of rutile so abundant in slates also seem to be rather readily affected, giving place to stouter crystals of the same mineral, or less commonly to anatase or brookite. Another early effect of metamorphism is the production of little flakes of brown mica (probably the haughtonite variety of biotite) from chloritoid substances, etc. With this there may be a crystallization of iron ores (magnetite or pyrites). In some cases a chlorite mineral or ottrelite is formed instead of the mica. In rocks rich in alumina chiastolite is produced concurrently with biotite.[2]

[1] Harker, Alfred, Petrology for students, Cambridge, 1895, pp. 264–265.

[2] For good colored figures see Teall, Pl. XXXIII, fig. 2, Fouque and Levy, Pl. III, fig. 1.

With advancing metamorphism graphitic spots and chiastolite crystals are lost, and the metamorphism begins to affect the whole body of the rock, the chief products formed being usually quartz and biotite. Of these the latter often has its flakes oriented in accordance with the original lamination or cleavage of the rock, and we have thus one type of mica-schist (Ger. Glimmerschiefer). These rocks may have no trace of the original clastic nature of the deposit, except perhaps some minute angular quartz grains. They sometimes shew a spotted character quite different from that mentioned above, and consisting in little ovoid spaces free, or relatively free, from the flakes of biotite which crowd the rest of the rock. Such spaces often shew distinctly crystalline properties, giving extinction parallel with their length, and in many cases, at least, they are ill-developed crystals of andalusite. When andalusite is better developed, it appears in clear crystal grains or in crystal plates enclosing other minerals; both forms are seen in the Skiddaw district, where andalusite-mica-schists have been extensively formed. Other minerals, such as white mica and little garnets, occur more locally."

The changes in mineral composition in the anamorphism of clays by contact metamorphism are shown in Fig. 11.

Chemical changes in contact metamorphism of shales and slates. The chemical changes involved in contact metamorphism of shale and slate are shown in Plate X.

Following is a list of references to the analyses platted:

1. Phyllite, Schneeberg, Saxony. Neues Jahrbuch, pt. 2, 1897, p. 156.
2. Shale, Arizona. Bull. 228, U. S. Geol. Survey, p. 348.
3. Upper Huronian (Virginia) slate, Mesabi district, Minnesota. Leith, C. K., Mon. 43, U. S. Geol. Survey, pp. 170 and 172.
4. Clay slate, Yaqui Gulch, Utah. Bull. 150, U. S. Geol. Survey, p. 342.

5a, 5b. Shale, Mt. Willard. Hawes, G. W., Albany granite and its contact phenomena: Am. Jour. Sci., vol. 21, 1881, pp. 7–8.

6. Shale, metamorphism at granite contact, Lerchenberg bei Zschorlau. Roth, J., Geologie, 3, p. 125.

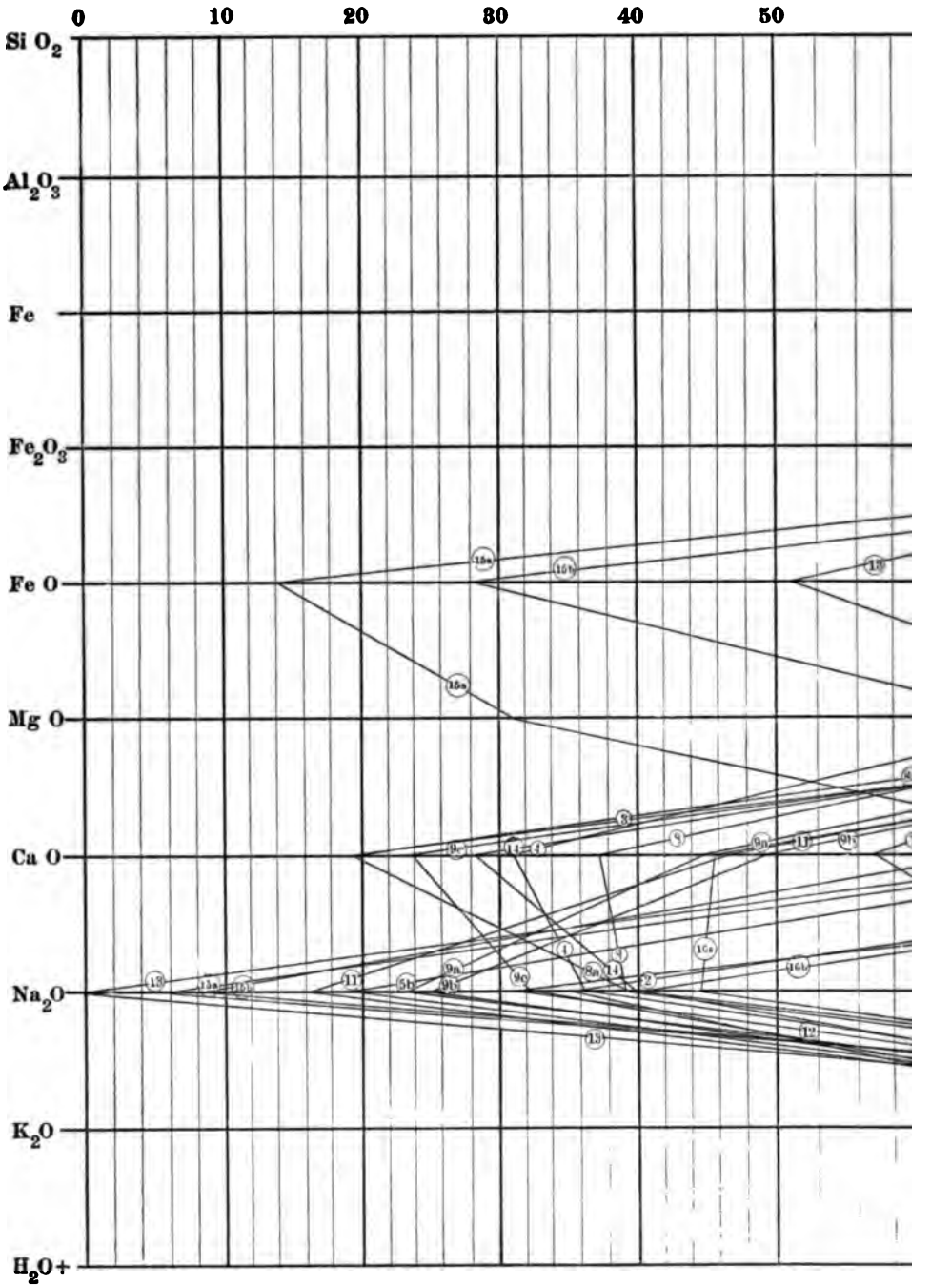
7a, 7b. Shale, Gabel. Roth, J., Geologie 3, p. 125, XII, 13, 14.

8, 8a. Shale, Allrode im Harz. Rosenbusch, H., Elemente der Gesteinslehre, 3d ed., 1910, p. 422.

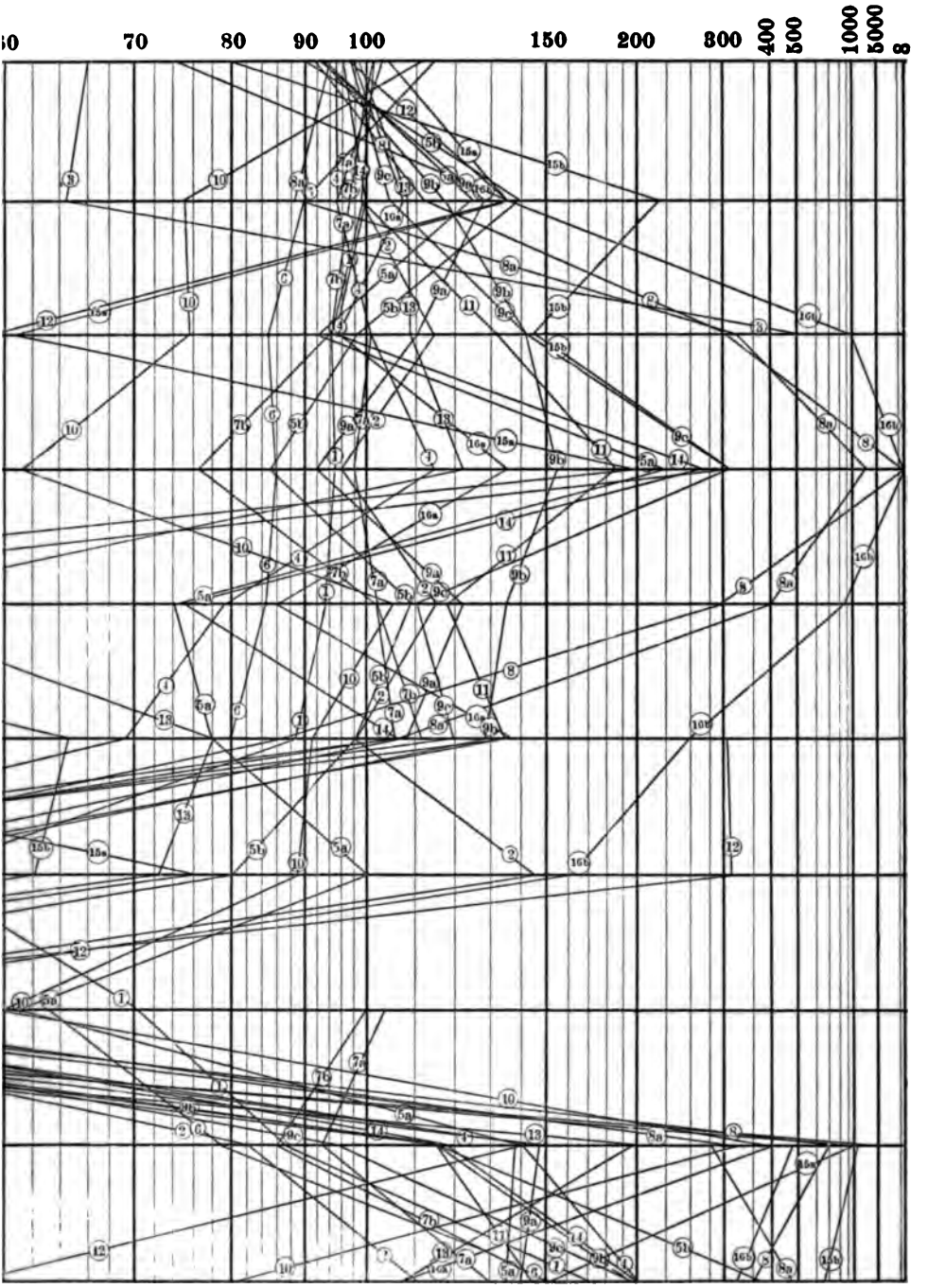
9a, 9b, 9c. Shale, metamorphism at diabase contact, Ruhrthal. Roth, J., Geologie 3, p. 147.

10. Shale, metamorphism at diabase contact, Mittelkopfes bei Hasselfelde. Roth, J., Geologie 3, p. 143, VII and 3.
11. Shale, metamorphism at diabase contact, Kuhlenberg bei Silbach. Roth, J., Geologie 3, p. 147. (Schenck, A., Verhandl. des na-

PLATE X



Representing the losses and gains of constituents in the contact metam
of analyses, numl



ism of shale and slate. Each broken line represents comparison of a pair
l for identification.

turhist. Vereins der preuss. Rheinl. und Westf. 1884, pp. 122–128.)

12. Shale, metamorphism at diabase contact, bei Weilburg an der Lahn. Roth, J., Geologie 3, p. 147. (Greim, Jahrb. Miner. 1888, I, 14.)
13. Shale, Lingolsbacher Thal. Roth, J., Geologie 3, p. 111.
14. Thonschiefer, Langenberge. Roth, J., Geologie 3, p. 104.
15a, 15b. Shale, metamorphism at diabase contact, Crystal Falls district, Michigan. Clements, J. M., Mon. 36, U. S. Geol. Survey, p. 210.
16a, 16b. Shale, metamorphism at diabase contact, Heinrichsburg bei Magdesprung, Harz. Rosenbusch, H., Elemente der Gesteinslehre, 3d ed., 1910, p. 422.

Clearly the tendency is for the loss of water, carbon dioxide, and oxygen, that is, the materials that are not required by the schist-making silicates, the usual introduction of soda, especially along contacts with basic rocks, and the occasional introduction of lime and potash. The characteristic introduction of soda is registered mineralogically by the development of albite as a porphyritic constituent. It seems difficult to avoid the conclusion that the soda introduced comes from the igneous rocks. Fluorine, boron, sulphur, and other substances of this nature may be locally introduced. The amount of water expelled from shales during contact metamorphism may be very great. Analyses furnished in connection with the contact phenomena of the Albany granite in shale at Mt. Willard [1] show that, on an average 52 liters of water were yielded from each cubic meter of shale in a zone extending 50 feet from the contact. The unaltered shale had a specific gravity of 2.85, and contained 4.09 per cent water, combined and mechanically enclosed; while the average altered shale in the 50 foot zone of metamorphism contained 2.27 per cent by weight of water. Thus, the loss of water due to contact metamorphism was equivalent to 1.82 per cent by weight of the original shale.

Fusion of shales and slates. Mineralogically, chemically and texturally the changes in shale or slate near igneous contacts are in the direction of producing an igneous rock. While one cannot deny that igneous rocks may have developed by the fusion of sediments of this type,—this change involving but little change in composition,—there are no clearly proved cases of any definite types of igneous rocks or any considerable quantities of igneous

[1] Hawes, G. W., The Albany granite, New Hampshire, and its contact phenomena: Am. Jour. Sci., vol. 21, 1881, p. 28.

rocks having been traced to this origin. The conclusion that igneous rocks have resulted from the melting of slates and muds must be based on *a priori* reasoning in our present state of knowledge of the subject. Attempts have been made, by comparing the composition of igneous rocks and slates near contacts, to show the probability of fusion, as, for instance, Bowen's [1] attempt to show that the granophyres or aplites connected with the Keweenawan gabbros of Ontario are in some cases the result of fusion of slates with which they are in contact. The granophyre is rich in albite and is reported as transitional into albite-rich adinole, a product of contact metamorphism of slates. Bayley [2] also regarded similar aplites or "red rocks" of Pigeon Point, Minnesota, as resulting from fusion of acidic sediments. Wright,[3] Collins,[4] and others who have studied the Keweenawan gabbros, regard the aplites as differentiates from the gabbro magmas on evidence that seems to be conclusive. Their composition shows eutectic relations with the associated basic igneous rocks. They are developed with the same composition in contact with sediments and igneous rocks of varying composition. They are developed with the same composition as dikes within gabbro masses far removed from contacts. Where the aplites come in contact with slate or quartzite along contacts it is usually with fairly definite boundaries.

Hobbs [5] has argued that lava magmas or "maculæ" result from fusion of shales on a large scale at a depth of about 11 miles or more below the surface. This argument is based: (1) on the similarity of composition between shales and igneous rocks (this correspondence is taken to offer an explanation for the limited range in composition of igneous rocks); (2) on the easy fusibility of shales; and (3) on the disturbance of equilibrium be-

[1] Bowen, Norman L., Diabase and granophyre of the Gowganda Lake District, Ontario: Jour. Geol., vol. 18, 1910, pp. 673–674.

[2] Bayley, W. S., The eruptive and sedimentary rocks on Pigeon Point, Minnesota, and their contact phenomena: Bull. 109, U. S. Geol. Survey, 1893, p. 117.

[3] Wright, F. E., The intrusive rocks of Mount Bohemia, Michigan: Ann. Rept. Geol. Survey Michigan for 1908, 1909, p. 393.

[4] Collins, W. H., The quartz diabases of Nipissing district, Ontario: Econ. Geol., vol. 5, 1910, pp. 538–550.

[5] Hobbs, Wm. Herbert, Some considerations concerning the place and origin of lava maculæ: Gerlands Beiträgen zur Geophysik, vol. XII, part 2, 1913, pp. 329–361.

tween temperature and pressure, allowing liquefaction of shales at points where overlying competent beds, like limestone, have folded and lifted the load from the shales. These localities are supposed to be distributed in zones which lie near the interior margin of mountain arcs formed of underturned and usually underthrust folds. The relatively high content of lime and soda in the igneous rocks as compared with the shales is accounted for by supposing that lime and soda have been assimilated during the rise of the magmas toward the surface through a process of overhead stoping. The lime is supposed to be assimilated from limestone. The source of the soda is not clear.

In the nature of the case, the conclusion that magmas develop in this way cannot be proved or disproved by field or laboratory evidence yet available. Adequate discussion of the subject involves a wide range of more or less theoretical conditions which are beyond the range of the more direct observations to which attention is principally given in this text. The general problem of subcrustal fusion is briefly discussed in Chapter IV of Part III.

Localization of contact metamorphism. In the vicinity of a large plutonic mass, at considerable depth, anamorphic effects, directly traceable to intrusion, may be traced for a very considerable distance, even measured in miles. Around smaller masses the influences are ordinarily measured by a few feet. Near surface flows the change may effect only a few inches or a fraction of an inch, presumably due to the rapid cooling of the extrusives. The wide variation in the range of effect of intrusions is only partly explained by depth and slow rate of cooling. It has been supposed that important causes of such variation were the presence or absence of distinctly mineralizing and altering solutions, given off from certain magmas or from certain parts of magmas. There are other local variations which are at present explainable. At igneous contacts and along veins traversed by thermal waters, the slates are more easily anamorphosed than the quartz sands, but less easily than the limestones.

SUMMARY OF MINERAL CHANGES IN THE SEVERAL STAGES OF ANAMORPHISM OF A CLAY

Fig. 11 represents the mineralogic changes, as calculated from average analyses, through the successive stages of anamorphism

of a clay. On the same diagram the mineralogic changes from igneous rock to clay are introduced for comparison.

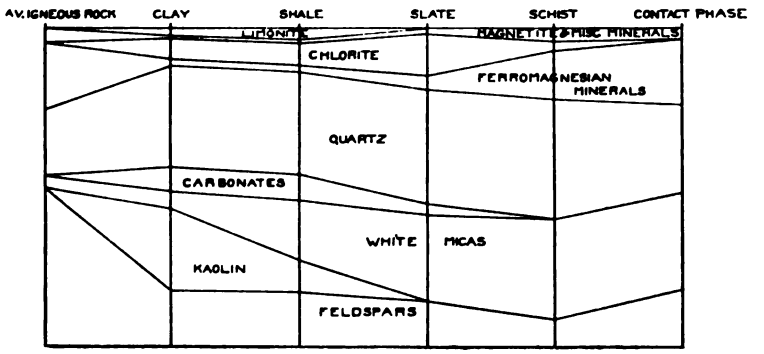


FIG. 11. Representing change in mineral composition in anamorphism of clays by cementation, dynamic, and contact metamorphism, as calculated from average analyses.

CHAPTER II

ANAMORPHISM OF SANDS (ARENACEOUS ROCKS OR PSAMMITES)

QUARTZ SANDS

Cementation of Quartz Sand to Sandstone and Quartzite

Quartz sands have their ultimate origin in the katamorphism and redistribution of granites and other igneous rocks containing free quartz. So far as they have passed through cycles of sedimentation they have their immediate origin in preëxisting psammites and other quartz-bearing sediments. The average mineralogical composition is given on page 76. When first deposited the pore space is likely to range from 20 to 40 per cent.

Cementation seems to begin, in some instances, almost as soon as the sand is deposited, whether under air or water. Water itself acts as a sort of cement by filling the openings and rendering the sand more firm and coherent. In this way certain beaches become hard and firm enough for automobile speedways. Under certain conditions lime, iron oxide, silica, bitumen, or clay may fill the spaces in the sand soon after it is deposited, producing a rock-like consistency. Sand dredges have brought up considerable blocks of recently deposited sand sufficiently cemented to hold together. Of particular interest is the possible early precipitation of quartz cement from colloidal silica solutions which are so often produced simultaneously with the fragmental products of rock decay. (See pp. 9–10 and 81.) Where deposited under air there are evidences of cementation, due to infiltration of lime, iron, silica, and other substances. Case hardening may result from surface evaporation of solutions carrying these substances.

The extent of cementation soon after deposition has not been fully investigated. It has been ordinarily assumed that cementation largely takes place later, during long periods of burial be-

neath the water table in the so-called belt of cementation. Unquestionably the process is important there, but it may have been preceded to an important extent by cementation near the surface, or when first deposited under water. There are comparatively few localities where uncemented sands extend down to the water table, and there give way to cemented sands.

Where the sand is partially cemented and breaks between the quartz grains it is called a *sandstone*. The common cements of sandstones are calcium (and magnesium) carbonate, quartz, and more or less hydrated iron oxide. One or another may dominate at different places. Calcite and quartz are in general more important than iron oxide. Iron oxide may color and coat sand grains without acting as a cement. It is of interest to note that the iron oxide and quartz are usually associated and that the presence of iron oxide usually means the absence of carbonate. In other words, they are often mutually exclusive. Cases are known of alterations along veins in a carbonate cemented sandstone where the new cement brought in is iron oxide and quartz which replaces the carbonate. There are cases also showing the reverse replacement.

In cementation of sand to sandstone there is a change in composition due to the introduction of iron oxide and carbonate. So far as quartz is introduced it does not change appreciably the percentage composition.

When the sand becomes so thoroughly cemented that it breaks through the clastic grains it is called a *quartzite*. The cement then is nearly all, if not entirely, quartz. The iron oxide and carbonate cements are ordinarily too weak to afford sufficient coherency to the rock, although there are cases of iron oxide cements in quartzite. As sandstones are often an intermediate stage in the formation of quartzite, it follows that in the latter the common carbonate and iron oxide cements have become replaced by quartz. Quartz may then be regarded as the permanent cement of sand, and carbonates and iron oxide as temporary cements in the sandstone stage of development. The composition of the quartzite does not differ from that of the sand but it does differ from that of the sandstone in showing less lime carbonate.

In general the older sands are more likely to have been cemented to quartz than the younger ones because they have

been longer subjected to the conditions of cementation. Thus there are usually quartzite and not sandstones in the earlier pre-Cambrian. But the fact that there are sands and sandstones as old as Paleozoic or late pre-Cambrian, and quartzites as late as Tertiary, shows that time is not the controlling factor in all cases. In these cases the presence or absence of appropriate solutions and conditions of precipitation seems to be the controlling factors.

The crystallization and texture of the cements show some variation. Lime carbonate may occur as coarsely crystalline calcite, as minutely crystalline calcite, or in amorphous, fibrous, or lamellar forms, of the types deposited from calcareous springs, streams, or in caverns.

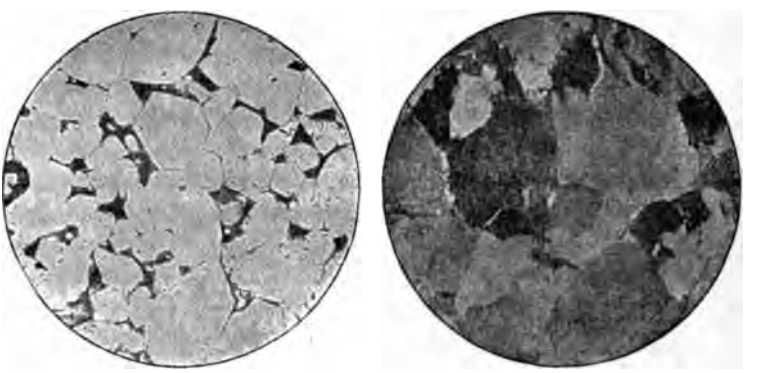


FIG. 12. Photomicrographs showing enlargement of quartz grains in cementation of sandstone and quartzite. A. Sandstone, ordinary light. B. Quartzite, crossed nichols.

The iron oxide also may be crystalline or lamellar, or amorphous.

Silica may be introduced in chalcedonic forms arranged regularly and peripherally with reference to the individual quartz grains, it may be introduced as a fine mosaic independent of the quartz grains, and very commonly it is introduced as *quartz enlargement*, a distinct growth of original quartz grains, the new part and the old part of the grain being in crystallographic continuity and making up a single crystal unit.

Quartz enlargement tends to develop the pyramidal crystal faces with the result that in intermediate stages of cementation, before the interstices are filled up, the rock is likely to exhibit innumer-

able brilliantly reflecting crystal faces of quartz. The continuation of growth fills the interstices and produces a massive quartzite. It is an interesting fact that many of the crystals meet one with another along plane surfaces corresponding in some cases to the terminal faces of one or more of the adjoining crystals. Under the microscope there is then to be seen a more or less clearly defined worn and rounded core of quartz, marked by a thin surface layer of ferric oxide or other impurities, with the new quartz added in optical continuity outside, usually, though not always, in clearer and more limpid form. The original sand grain, after being torn from its parent igneous rock, may have undergone a long and varied history through repeated erosion cycles, but when silica-bearing solutions came through the rock, these grains were able to rejuvenate themselves by growth. This is an interesting aspect of the wonderful power of crystallization, which, after interruptions of perhaps millions of years, exerts the same power as when the original crystal formed.

Quartzites may become so thoroughly recrystallized that all traces of original grains are lost, the rock being a dense mass of coarsely crystalline quartz. This may result from a continuation of the process of cementation. Complete gradation may be seen between ordinary cementation and such extreme recrystallization. Often such recrystallization has taken place under contact metamorphism, or in places of extreme deformation, but it is not always possible to prove these agencies.

Source of cementing materials. The cementing materials of the sands are in part derived from outside sources and in part derived from the sands themselves. Since the sands afford opportunity for a comparatively free flow of water, cementing materials may be carried long distances. For example, mineral matter taken into solution on an erosion surface may enter an artesian system and be carried underground for many miles, to be ultimately deposited near the point of escape by reduction in pressure, by mingling with other solutions, or by interaction with the constituents of the sand.

Silica cement may obviously be derived by solution of sand near the surface and deposition in the same sand below, or in arid regions upward moving waters may dissolve silica and deposit it at the surface by evaporation. Silica may also be intro-

duced from outside sources from the weathering of other formations. Its principal source is in the destruction of silicates by weathering.

Iron may be carried down as ferrous iron from overlying carbonate formations and be oxidized on meeting freely flowing oxidizing solutions in the sandstone. Also iron oxide cement may in small part be derived from oxidization of ferromagnesian fragments of the sand.

Lime may obviously be introduced from outside sources, as it is the most abundant constituent of most natural surface waters. Calcareous layers in a sand may be distributed as cement to neighboring parts of the same formation.

So far as unaltered silicates are present as constituents of a sand they may be altered and contribute silica and salts of the bases as cement.

So far as clay is present in the sand it may absorb potash by abstraction from circulating waters.

Dynamic Metamorphism of Quartz Sand Rocks

In some cases sandstones or quartzites which have undergone dynamic metamorphism are found to consist of masses of coarsely recrystallized quartz, showing no secondary development of minerals or parallel structures. Dynamic metamorphism is usually evidenced by a schistose structure, due to the slicing and granulation of original quartz grains, or to the recrystallization of quartz in bands, or to some combination of the two processes. Where sliced, the quartz fragments may be elongated in the plane of schistosity. Where recrystallized, the individual quartz crystals are not elongated to any considerable extent; they have about the dimensions determined by their normal habit. They do, however, lie in distinct bands which gives an impression of flattening and elongation. The quartz individuals are not crystallographically oriented. Evidences of recrystallization are the absence of strain effects in the quartz, the absence of angular fragments, the perfect joining of the adjacent particles without interstitial material, the tendency toward uniformity in size, and the frequent coarse grain indicating some integrating process.[1] Where there are sufficient impurities in the original

[1] Leith, C. K., Rock cleavage: Bull. 239, U. S. Geol. Survey, 1905, pp. 70–71.

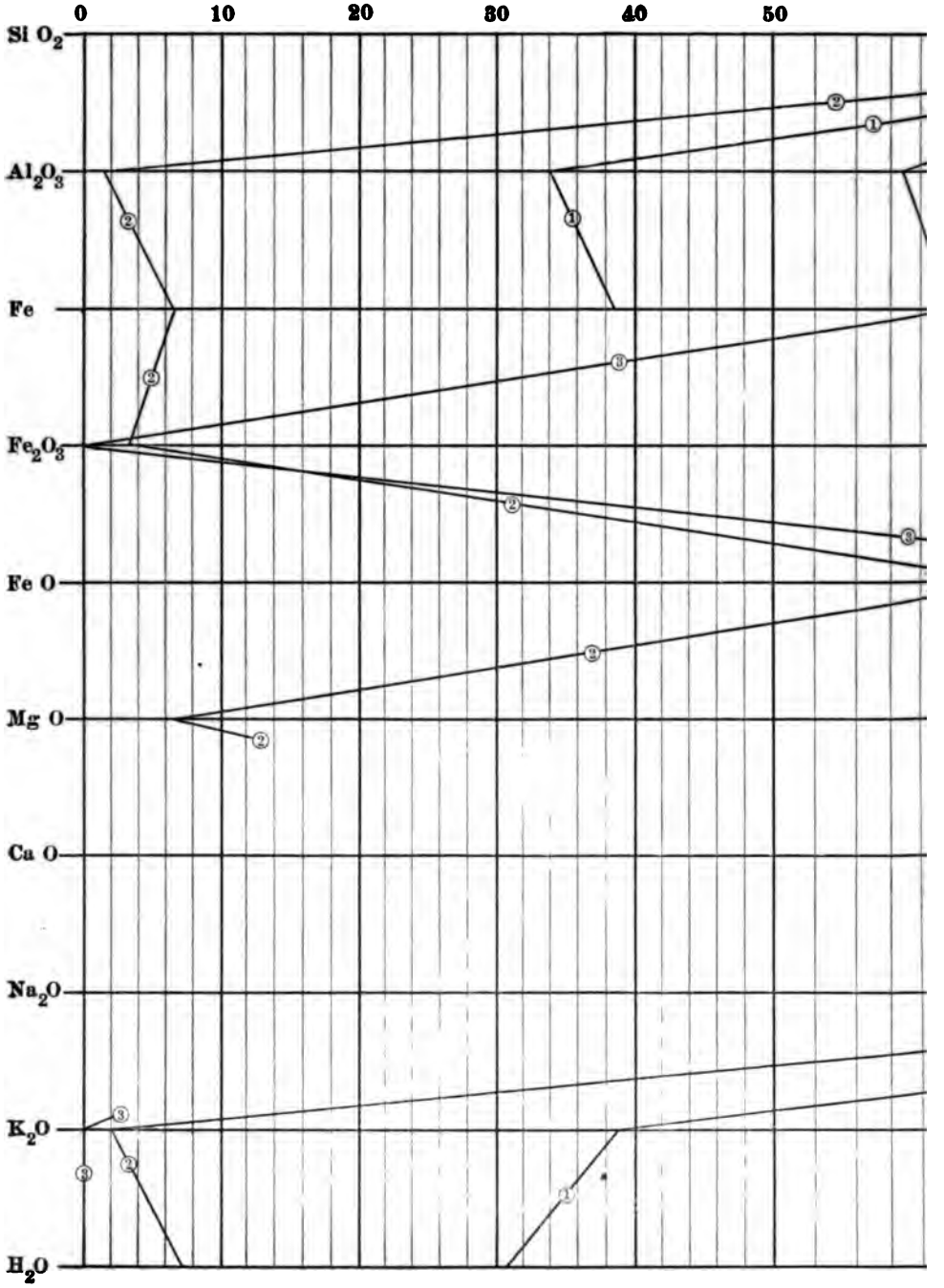
sand, mica (usually sericite) is likely to develop under dynamic metamorphism, producing a mica-quartz-schist. When the rock is cleaved it breaks along the mica bands—in fact along the mica mineral cleavage—for both surfaces of a cleaved rock are those of mica mineral cleavage. In extreme cases phenocrysts of garnet, staurolite, and other anhydrous silicates may develop with crystal outlines, usually without parallel arrangement, after the schistosity is produced, but the important developments of this kind are more characteristic of the quartz-feldspar-ferromagnesian sands.

Dynamic metamorphism may or may not change the composition of the quartzite. Where originally composed entirely of quartz, the sliced, sheared, or recrystallized end-products may be also entirely quartz. Where a secondary silicate, principally mica, has developed, this may involve merely a recrystallization of the impurities in the rock itself. There is quantitative evidence that in some cases the process involves considerable loss of quartz. Sericitic schists may be observed in shear zones, both parallel to and crossing the bedding of the quartzite. In an extreme case of shearing, sericite may develop almost to the exclusion of quartz. The conditions in these shear zones would seem to require diminution, rather than increase, in volume, thus favoring elimination rather than introduction of materials. That these shear zones were not originally of different composition than the remainder of the quartzites is shown by the fact that they cross the beds, thus including within the shear zone various beds which may have had different original compositions. Properly taken representative samples of these sericitic shear zones have been analzyed and compared with analyses of similar samples of the original quartzite of the beds abutting the shear zones. It appears that the relative proportions of material other than silica are the same in the original quartzite and in the sericite schist, but that the proportion of quartz has been greatly diminished. In other words, the composition of the sericite schist is the same as that of quartzite, minus quartz. (See. Pl. XI.)

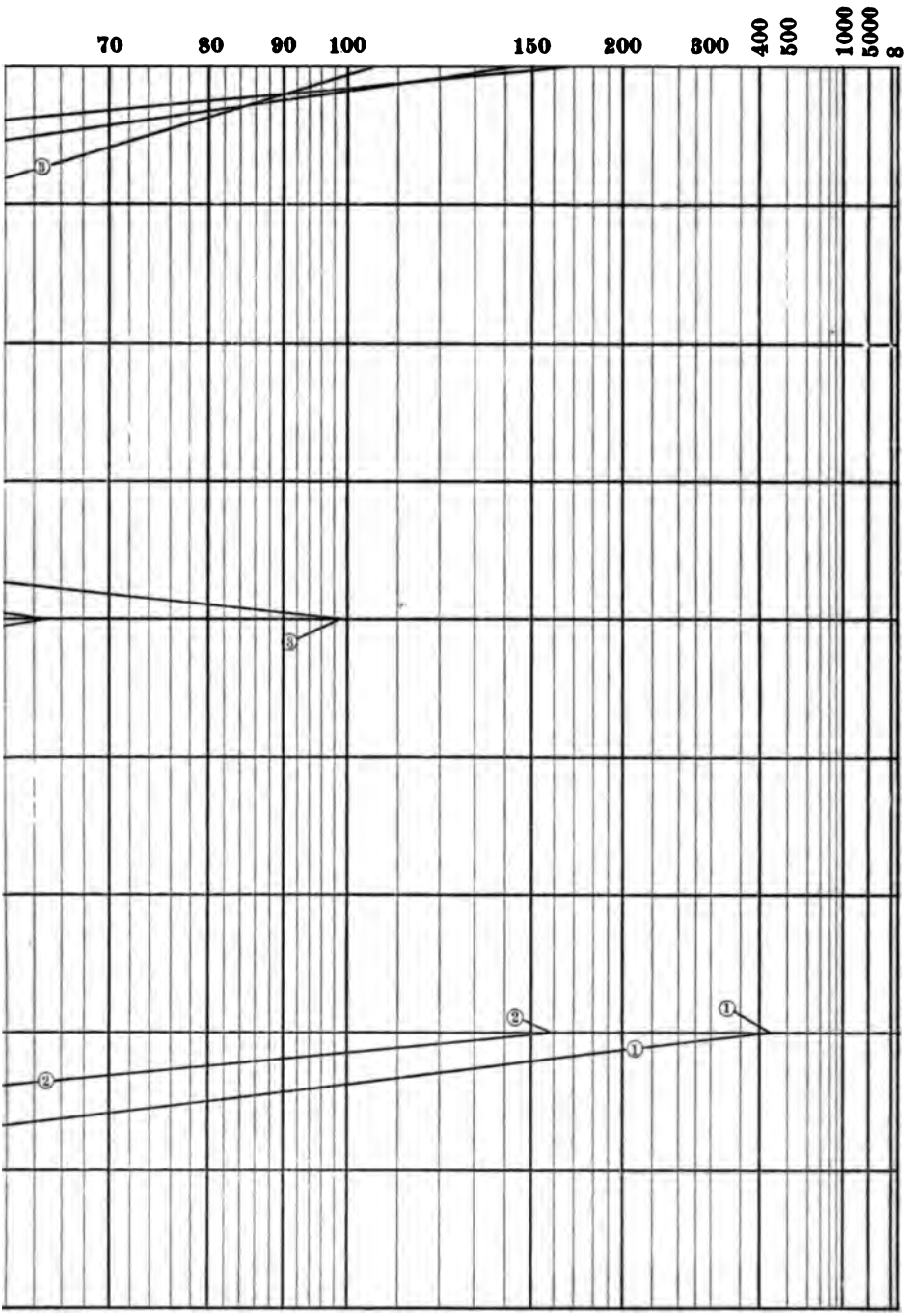
Following is a list of references to the analyses platted:

1. Waterloo quartzite, Waterloo, Wisconsin. Warner, J. H., The Waterloo quartzite area of Wisconsin. Unpublished bachelor thesis, Univ. of Wis., 1904.

PLATE XI



Representing losses and gains of constituents in dynamic metamorphism
a pair of analyses, nu



uartzite to sericite schist. Each broken line represents comparison of
d for identification.

2. Mesnard quartzite, Marquette Range. Mon. 52, U. S. Geol. Survey, p. 257.
3. Mesnard quartzite, Marquette District. Leith, C. K., Unpublished analyses.

Introduction from outside sources of part or all of the constituents of the sericite, without elimination of quartz, would involve an increase of volume, and would further require the introduction of substances from without in proportions almost identical with that of the composition of the impurities in the quartzite, which would be a remarkable coincidence, In the discussion of the dynamic metamorphism of other rocks, it will appear that the elimination of water, carbon dioxide, oxygen, and lime, has been proved in certain cases. It is our view that silica may be eliminated in cases of the sort just described. We see no adequate reason for accepting the elimination of certain constituents, and denying, because of supposed insolubility, the elimination of others. By reasoning from considerations of this sort, we have been led to the formulation of a principle, discussed on page 203, that in the development of schists by dynamic metamorphism there may be elimination of any of the constituents of the rock which are present in excess of the amounts required for the production of the common schist-making minerals. In the quartzite, quartz is present in excess of such requirements.

Under extreme conditions of anamorphism a quartz-mica-gneiss may develop from a quartzite with growth of new feldspar and coarse recrystallization. These are perhaps most likely to develop from impure feldspathic quartzites, but it is not impossible, in some cases, that they represent an extreme phase of anamorphism of mica-quartz-schists, in which there was originally but little feldspar. The gneiss is relatively anhydrous.

Thermal and Contact Metamorphism of Quartz Sand Rocks

Contact metamorphism may cause cementation, coarse recrystallization, or schistosity, which in some cases are distinguished with difficulty from the ordinary results of cementation and dynamic metamorphism. An illustration is the Rib Bill quartzite of Wisconsin, which has been intruded by granite and

much deformed, and now consists of a thoroughly recrystallized coarse, clear quartz mosaic without trace of original clastic structure, and, for the most part, without schistosity. How much of this is due to dynamic metamorphism, and how much to contact metamorphism, has not been fully ascertained. · There is, however, a general type of metamorphism which can be easily traced to contact action, involving the development of porphyritic minerals, particularly garnet, sometimes tourmaline, especially where the rock has some impurities in it, and the development of a spotted texture, marked by the segregation of amphibole, chlorite, calcite, epidote, and other minerals. The spots are often encircled by light-colored rings, suggesting the segregation of the original ferrous iron of the quartzite, and that the alteration involves simply rearrangement of original constituents and requires little change in composition of the rock as a whole. A secondary development of orthoclase feldspar, giving the rock a reddish color, has been noted at some contacts. This has been taken in some cases as a suggestion that certain acid red rocks of igneous type are formed from fusion of quartzite. (See pp. 117–118.) The total chemical change in the quartzite in these cases has not been determined. One pair of analyses of spotted quartzites from the Pigeon Point district of Minnesota seems to show an increase in potassa and especially in soda, but this is not enough basis for generalization. The introduction of materials from magmatic sources into the quartzite is principally confined to alkalies.

Volume Changes in Cementation and Anamorphism of Quartz Sand

The volume changes in cementation and anamorphism of quartz sands seem to be slight in the early stages.

The question of volume change during cementation of a sand is one which has not yet been adequately investigated. Presumably during the early stages there is some tendency towards decrease in volume by the settling of the formation and the elimination of porosity and water. The filling of the voids by cements necessarily eliminates porosity. The crystallization of the infiltrated cements may actually tend to increase volume, because the substances crystallizing from solution are known experimentally

to exert pressure.[1] Becker [2] has attempted to explain the widening of fissures in the Mother Lode district of California by the crystallizing of quartz. The same explanation has been used to explain the separation of fragments in a cemented breccia. Microscopical examination of quartzites shows that a surprisingly large proportion of the original quartz grains are not in actual contact, but are completely separated by cements, suggesting the possibility of the fragmental particles having been forced apart during the recrystallization of the quartz.

The average density of the minerals of the rock is not changed, except where calcite and iron oxide come in as cements. These are heavier than the original quartz grains.

When the quartz schist develops there is probably reduction of volume, due to the elimination of silica, to the recrystallization of impurities into silicates, and to the reduction of any remaining pore space. The chemical changes in the early stages of the process are merely those of addition; in the later stages of dynamic anamorphism they are principally those of elimination of part of the substances in excess of those required to make up the micas—in other words, principally the quartz. Near igneous contacts some substances are apparently introduced.

Locus of Cementation and Anamorphism of Quartz Sand

The above changes are described with only incidental reference to location of the change in any particular zone—they may in fact occur under a variety of conditions. Cementation begins with deposition of the sand grains. Case-hardening of sandstone is cementation going on directly at the surface, due to evaporation of solutions from within the sandstone. In general the evidences of cementation are likely to be more conspicuous beneath the surface, and especially below the water table. Schistose structures indicate zones of excessive shear, as, for instance, between strong formations as along fault planes. These schistose shear zones may develop at almost any depth, but usually indicate a considerable load. Contact effects may develop at almost any depth but are

[1] Becker, G. F., and Day, A. L., Linear force of growing crystals: Proc. Wash. Acad. Sci., vol. 7, 1905, pp. 283–288.

[2] Loc. cit., p. 287.

more notable near large plutonic masses which have cooled slowly. Quartz sand rocks because of their mechanical strength and chemical stability, yield much less readily to anamorphic agencies than do shales and limestones.

QUARTZ-FELDSPAR-FERROMAGNESIAN SANDS

Cementation of Quartz-Feldspar-Ferromagnesian Sands

The account of the cementation of quartz sands on a preceding page may be applied with little modification to the cementation of quartz-feldspar-ferromagnesian sands. By cementation quartz-feldspar sand becomes arkose and quartz-feldspar-ferromagnesian sand becomes graywacke. Being less simple mineralogically, the process of cementation does not stand out so clearly, but there is no reason to believe that there is anything essentially different in the process from that already described. The presence of feldspar and ferromagnesian minerals offers opportunity for metasomatic alterations which aid cementation. Secondary growths or enlargements of both feldspar and hornblende have been noted. The effect of these on cementation is presumably the same as is that of the enlargement of the quartz, although of course on a very much smaller scale. There is no case known where enlargement of these minerals alone has progressed to a sufficient extent to cement the rock. The grains of feldspar and ferromagnesian minerals undoubtedly undergo katamorphic alterations to some extent subsequent to deposition. These alterations characteristically involve increase in volume and consequently their effect on the rock as a whole is consolidation. Silica and soluble salts yielded to solution by these alterations are available for cementation through crystal enlargement or cavity filling elsewhere in the rock. So far as cementation is affected by increase in volume of individual grains through hydration it is in a sense cementation by addition of water. So far as these minerals are enlarged, or so far as new silicates are developed by metasomatic alterations, it has been a question as to what extent they are due to ordinary conditions of cementation, and to what extent they involve the agency of hot solutions from igneous sources.

Dynamic Metamorphism of Quartz-Feldspar-Ferromagnesian Sand Rocks

What has been said about the dynamic metamorphism of quartz sands applies, with little modification, to the quartz-feldspar-ferromagnesian sands. Containing as they do, more alumina and bases, they favor the more abundant development of secondary silicates during dynamic metamorphism. From such rocks mica schists, mica gneisses, hornblende schists, hornblende gneisses, and other platy rocks of this type develop. Porphyritic developments of garnet, staurolite and other anhydrous silicates, without parallel arrangement, are common. The dynamic metamorphism of these rocks has many features in common with the dynamic metamorphism of shales and slates described on pages 109–111. There has been but little quantitative investigation of these changes. We are unable to present analyses expressing this alteration.

Contact Metamorphism of Quartz-Feldspar-Ferromagnesian Sand Rocks

The statement on the contact metamorphism of the quartz sand group needs little modification to make it applicable to the quartz-feldspar-ferromagnesian sands, except that the changes in this group are more conspicuous. The presence of the feldspar and ferromagnesian minerals affords material for recrystallization into porphyritic minerals and spots to a larger extent than the quartzite. Coarse clear grains of feldspar, biotite, pyroxenes, and amphiboles are common developments of extreme anamorphism. Along igneous contacts feldspar (albite) is likely to be especially abundant, and may have a porphyritic development quite independent of schistose arrangement, obviously developed later than the schistosity, under conditions which allowed growth in directions independent of pressures. Such an arrangement is beautifully illustrated in the Hoosac formation in the Green Mountains of Massachusetts.[1] The contact metamorphism of this group also has many features in common with that of the shales and slates described on pages 114–117. No quantitative data on these alterations are available.

[1] Pumpelly, Raphael, Wolff, J. E., and Dale, T. Nelson, Geology of the Green Mountains in Massachusetts: Mon. 23, U. S. Geol. Survey, 1894, pp. 59–63.

Gradation of Quartz-Feldspar-Ferromagnesian Sands into Clays

As the quartz-feldspar-ferromagnesian sands become fine-grained and take on kaolinic constituents, they approach the clays or muds in composition and share in the kinds of anamorphism which have been described for the clays.

CHANGE TOWARD IGNEOUS TYPE UNDER CONTACT AND DYNAMIC METAMORPHISM OF SANDS

Texturally, mineralogically, and probably chemically, the changes above described in dynamic and contact metamorphism of the sand group are in the general direction of producing igneous rocks, but except locally at igneous contacts, the ultimate products do not become igneous rocks, but rather schists and crystalline rocks differing in certain essential particulars from igneous rocks. (See pp. 199–201.)

The change under dynamic metamorphism is the production of silicates, particularly mica, and the elimination of quartz. Under contact metamorphism the same change goes on but there is in addition the development of other silicates, such as albite, garnet, epidote, etc., which may represent, in part, the materials from igneous sources. Silicates under contact metamorphism become more numerous, more complex, and more varied, tending thus to produce the variety which is found in the igneous rocks.

Direct evidence of fusion of the quartz sand into igneous rocks on any large scale is not yet available. Partial gradations between the two types of rocks may be traced in some cases, but it has been difficult to prove satisfactorily that any particular type of igneous rock, or any considerable body of igneous rock, is a direct result of fusion of the quartz-sand group. Various cases have been cited of intrusion of basic igneous rocks into acid sediments with the development of an acid red rock or aplitic phase at or near the contact, very suggestive of fusion of the acid sediments and possibly accompanied by some sort of a differentiation or density stratification. (See pp. 117–119.) Analysis of the evidence in these cases leads to doubt as to the validity of this conclusion. For instance, there is clear evidence that the existence of red rocks and aplite associated with the basic Keweenawan

intrusions of the Lake Superior country and Ontario are not dependent on the nature of the intruded rock; there are many contacts in the basic igneous rocks and the quartzite in which this type does not develop. An eutectic relationship has been developed between the acid igneous after-effects of this type, on the one hand, and the basic rocks on the other, and all gradations between the two identified. It is not our purpose to assert that fusion of sediments may not go on on a large scale and that certain types of igneous rocks may not result from this fusion. We wish only to emphasize the point that the actual evidence is limited and that there is a great gap between the sediments on the one hand and the igneous rocks on the other, which, in the nature of the case, it is difficult to bridge by inductive reasoning. The origin of acid igneous rocks of possible sedimentary origin is so involved with the questions of origin of igneous rocks in general that a definite conclusion as to the origin does not seem to be possible with present means of attack on the problem. The matter of fusion of sediments is considered, in its larger aspects, on another page. (See pp. 268 et seq.)

CHAPTER III

ANAMORPHISM OF THE CARBONATES

GENERAL ACCOUNT

The anamorphism of calcium carbonates, calcium-magnesium carbonates, and iron carbonates proceeds along closely similar lines. To discuss the anamorphism of each of these carbonates independently would involve considerable duplication. For convenience and brevity the discussion will be devoted mainly to the calcium carbonates, with incidental reference to the magnesium and iron carbonates, where the anamorphism of these rocks produces distinctively different results.

Calcium carbonate formations may have passed through few or many generations of solution and sedimentation but their ultimate origin is in the destruction of basic silicates of the igneous rocks. The breaking down of igneous rocks usually liberates silica, which is in part deposited with the limestone as chert in bands and scattered through the calcium carbonate layers.

From the moment of deposition there commence the processes of settling, consolidation, cementation, and recrystallization, which result in the formation of a limestone. The mechanical breaking down of coral reefs and the settling of fragments in openings is a phase of the process. Pore space is greatly reduced, the small particles coalesce into larger crystal units, water is expelled, and the mass becomes integrated. Marl and calcareous oozes become limestone. Complete gradation may be observed from exceedingly fine-grained limestones to crystalline limestone of coarser texture.

The continuation of these processes of integration means further elimination of pore space and water and coarser crystallization. The product is marble, in which the individual crystal units may be several hundred or several thousand times the size of those in limestone. The mass may or may not become

schistose. Usually it does not, due probably to the fact that calcite is not adapted in form to the lamellar arrangement of schists, and due also to the fact that it recrystallizes very readily, tending to destroy evidences of schistosity which may have at one time existed.

The next stage in the anamorphism of calcium carbonate rocks is the development of silicates, often with lamellar or porphyritic arrangement. As anamorphism proceeds, the silicates become more abundant and finally the original limestone may be represented entirely by the silicate rock of the type of amphibolite or schist, or of the fine-grained silicated phases characteristic of some igneous contacts.

During any of the stages of anamorphism of a carbonate, from the unconsolidated carbonate to the marble or silicate, there may be silicification by the introduction of chalcedonic quartz, replacing the carbonate. There is some evidence that this process begins almost from the moment of deposition, for silicified fossils are found completely imbedded in carbonate. The intimate association of chert and limestone cannot be always satisfactorily explained as due to introduction of silica from without after the rock has been indurated, but seems in some cases to be due to primary precipitation with the carbonate. Underground meteoric solutions accomplish secondary silicification along cracks, bedding planes, and other openings. Hot waters along or near igneous contacts may extensively silicify carbonate formations. There is some evidence that this process is more or less selective in that certain beds seem to be more easily attacked than others, and that certain shells are silicified while other parts of the rock mass are not.

The entire process of integration from unconsolidated calcium carbonate deposits to the silicated or silicified limestone or marbles presents a continuous series of alterations, one merging into the other. The earlier phases of the change may go on under water or at moderate depths in the earth; the later stages are usually associated with rock flowage or deep burial, or at igneous contacts where temperature and pressure are high. Limestones yield to rock flowage more readily than sands, but less readily than shales. They yield to contact metamorphism more readily than shales and sands. Carbonates may undergo anamorphism

under the same conditions that may katamorphose an igneous rock.

The earlier processes of cementation are comparatively simple. They have been frequently observed and described. The later stages under dynamic and contact metamorphism are not so simple and present a number of problems, some of which have important bearing on problems of economic geology, for the anamorphic phases of limestones occur frequently along igneous contacts which are also the loci of ore deposits.

DYNAMIC METAMORPHISM OF LIMESTONE

The commonest result of dynamic metamorphism of limestone is crystalline marble. Many instances might be cited of limestones and marbles in the pre-Cambrian which have been folded and sheared between beds of gneiss or sediments and still retain a massive texture, although the associated beds may be highly schistose or gneissic. The marble itself may have been schistose at one time, but this texture has been completely obliterated by recrystallization.

Exceptionally the limestone or marble may be schistose, without secondary development of silicates. In some cases the schistosity seems to be due to a granulation of originally larger fragments of calcite, and a strewing out of the fragments into bands, as in the case of the limestone from Talladega Mountain in the Ocoee series of the Southern Appalachians. In other cases there has been a minute parallel slicing of the original mass of rock, giving a structure which might be called fracture cleavage. Here there has been no recrystallization to obliterate the structure. The rarity of schistose marbles or limestones as compared with schistose phases of almost any other rock is probably due to the remarkable readiness with which calcite recrystallizes. The coarsely crystalline character of marbles which have undergone rock flowage is itself evidence of the importance of recrystallization in this rock. Where the limestone contains sufficient impurities, amphiboles or mica, or talc, or chlorite may develop with parallel arrangement, giving the rock a crude schistosity, although the calcite itself may not contribute to the production of the schistosity. In extreme cases, practically all of the carbonate may be eliminated, as in certain talc schists and possibly

certain amphibolites. This has meant elimination of carbon dioxide and lime with consequent great loss in volume and retention of all other substances in the rock. Johnston [1] found that ordinary grinding of calcite eliminated some carbon dioxide. There is little requirement for or evidence of the introduction of substances from without, but such additions may occur. Where such work develops along a shear zone, requiring diminution in volume, it is easier to conceive of the change being accompanied by elimination than by substitution or addition.

Porphyritic development of anhydrous silicates, like garnet, diopside and tremolite, independent of the schistosity are common developments in dynamically metamorphosed marbles. These are clearly due to later recrystallization of the marble and its impurities after movement has ceased, comparable to the late porphyritic growths in mica and chlorite schists and slates. (See p. 180.)

THERMAL AND CONTACT METAMORPHISM OF LIMESTONE

Limestones are more readily altered under conditions of thermal and contact metamorphism than other sediments. The contact phases are favored spots for the accumulation of certain ores.

At the surface in the vicinity of volcanoes, and in fissures along which thermal waters are traveling, limestones may be extensively dissolved, silicified, dolomitized, sericitized, and chloritized. Ordinarily the chloritization and sericitization take place below the surface. Locally, siderite, magnesite, and fluorite may develop. These processes are essentially anamorphic in their nature, though near and at the surface there may be solution as in katamorphism.

In contact with surface flows or with small dikes the limestone may not be greatly affected, but it may become recrystallized or silicified. The quartz takes the form of chalcedony or chert, which may be iron-stained, producing a jasper.

Along deep veins the action of hot waters has been to replace the limestone by epidote, garnet, pyroxene, amphibole, biotite, and sometimes tourmaline. This alteration is substantially the same as that described under contact metamorphism in the following paragraphs.

[1] Johnston, John, and Niggli, Paul, The general principles underlying metamorphic processes: Jour. Geol., vol. 21, 1913, p. 614.

Where intruded by larger igneous masses, some distance below the surface, the limestone is likely to be much more extensively altered. It may become crystalline marble with minor quantities of secondary silicates such as amphiboles, pyroxenes, garnets, etc. It may be replaced entirely by coarsely crystalline silicates of this type. It may be a fine-grained brick-like silicate mass without calcite.

Plate XII compares analyses of contact phases of limestone with original phases.

Following is a list of references to the analyses platted:

1. Homestake limestone, Iron Springs, Utah. Leith, C. K., and Harder, E. C., Bull. 338, U. S. Geol. Survey, pp. 26–27.
2. Morenci limestone. Kemp, J. F., Econ. Geol., vol. 2, 1907, p. 7.
3. Limestone, Bingham district, Utah. Boutwell, J. M., Prof. Paper 38, U. S. Geol. Survey, p. 191.
4. Naco limestone, Bisbee district. Ransome, F. L., Prof. Paper 21, U. S. Geol. Survey, p. 148.
5. Limestone, White Knob, Idaho. Kemp, J. F., Econ. Geol., vol. 2, 1907, p. 6.
6. Jurassic limestone, Chanaicillo, Chile. Natur Marburg, 1870, p. 26.
7. Limestone, San José, Mexico. Kemp, J. F., Econ. Geol., vol. 2, 1907, pp. 6–7.
8. Limestone, Clifton-Morenci district. Lindgren, W., Prof. Paper 43, U. S. Geol. Survey, p. 71.
9. Limestone, Bingham, Utah. Boutwell, J. M., Prof. Paper 38, U. S. Geol. Survey, p. 191.
10. Limestone, San José' Mexico. Kemp, J. F., Econ. Geol., vol. 2, 1907, pp. 6–7.
11. Limestone, Hastings, Ontario. Adams, F. D., Jour. Geol., vol. 17, No. 1, 1909, p. 11.

12, 13. Limestone, San Francisco district, Utah. Butler, B. S., Prof. Paper 80, U. S. Geol. Survey, p. 84.

14a, 14b. Limestone altered to amphibolite. Adams, F. D., and Barlow, A. E., Geology of the Haliburton and Bancroft areas, Province of Ontario: Memoir No. 6, Dept. of Mines, Canada, p. 104.

This plate gives an idea of the net change accomplished at contacts. It does not, however, tell much about the processes which have produced the change.

CONTACT AND DYNAMIC METAMORPHISM OF IRON CARBONATES (SIDERITE)

The early stages of cementation and induration of iron carbonates proceed along the lines already described for the calcium carbonates. Under dynamic action and at igneous contacts, the processes are also approximately similar to those produced by the anamorphism of calcium carbonates, but distinctive minerals, high in iron, are developed, producing a product which differs in certain features from that of the anamorphism of calcium and magnesium carbonates.

The anamorphism of the iron carbonates has been studied quantitatively both in detail and on a large scale in connection with iron formations of the Lake Superior region. Near igneous contacts, and where the carbonate has been considerably sheared, there is extensive development of iron amphiboles and magnetite, and locally garnet and chloritoid. The iron-rich amphiboles are formed by a combination of the ferrous iron (and other bases present as impurities) with silica which is abundantly present in the form of chert. These often develop without parallel arrangement, in an interlacing network, particularly along contacts of the carbonate and chert layers. Part of the iron is partially oxidized to form magnetite, which is a characteristic associate of the amphibole. The excess of chert present recrystallizes into coarse quartz mosaic. The characteristic banded structure of the cherty iron carbonates is more or less retained in the anamorphic rock, being marked by alternating bands of quartz, amphibole and magnetite. These rocks are called amphibole-magnetite rocks or schists.

The chemical change is principally the elimination of carbon dioxide, the partial oxidation of the iron, and the combination of part of the iron with the silica. There is no evidence of the introduction of materials from without during this change, as the composition may be easily accounted for by the elimination of carbon dioxide from the primary carbonates. Analyses of these anamorphic products of cherty iron carbonate have been made on a large scale in the Lake Superior region, and compared with the primary carbonates. They indicate that the ratio of silica to iron has remained substantially unchanged.

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SOURCE OF SILICATES IN THERMAL AND CONTACT METAMORPHISM OF CARBONATES

Difference in opinion has arisen as to the relative importance of magmatic contribution as a source for the silicates, or the recrystallization of impurities in the limestone, with elimination of carbon dioxide and possibly lime. Both have been proved in specific instances. The question is one of relative importance of the two processes.[1]

All would agree that there has been replacement to some extent from outside sources, presumably magmatic, for frequently these contact phases contain both distinct veins and irregular masses of quartz or sulphides or magnetite, or iron garnets, which could not be regarded as residual from the limestone because of their composition. Soda is clearly introduced. One of the potent reasons for believing the silicated phase to represent essentially replacement is the fact that this replacement would maintain the volume, and field evidence, such as retention of bedding, often indicates that volume has been essentially unchanged during alteration. On the hypothesis that the contact phase is the result of elimination of calcium carbonate, a very large reduction of volume is necessary—perhaps up to 80 or 90 per cent. Many geologists have regarded so large a reduction as inherently improbable, and unsupported by field evidence. It has been suggested also that the similarity of silicates and other contact minerals along contacts of different kinds of sediments and both acid and basic rocks, is in itself evidence of introduction of substances from without.[2]

Likewise, there is substantial agreement that elimination of carbon dioxide and lime and recrystallization of the remaining constituents have also played a part in forming silicate contact zones, but there is difference of opinion as to the importance of this process. Evidence of recrystallization is as follows:

[1] For summary of arguments on the two sides of this question see: Lindgren, W., The origin of the "garnet zones" and associated ore deposits: Bull. 90, Am. Inst. Min. Engrs., June, 1914, pp. 949–956; Econ. Geol., vol. 9, 1914, pp. 283–292.—Leith, C. K., Recrystallization of limestone at igneous contacts: Bull. 90, Am. Inst. Min. Engrs., June, 1914, pp. 1129-1134; Econ. Geol., vol. 9, 1914, pp. 292-299.—Uglow, W. L., A review of the existing hypotheses on the origin of secondary silicate zones at the contacts of intrusives with limestones: Econ. Geol., vol. 8, 1913, pp. 19-50 and 215-234.

[2] Spurr, J. E., and Garrey, G. H., Ore deposits of the Velardena district, Mexico: Econ. Geol., vol. 3, 1908, pp. 688–725.

(1) The development of silicates from the calcium or magnesium carbonates involves the elimination of all the carbon dioxide and some of the lime and magnesia, with recrystallization into silicates of part of the lime and magnesia, together with other impurities which may be present, such as silica, iron, kaolin, and other substances. In certain districts the composition of part of the silicate zone (usually the outer part) corresponds approximately to the composition of the original carbonate rock, minus carbon dioxide and a part of the lime and magnesia. No analyzed samples have shown exact correspondence. It would be difficult to find exact correspondence because of later replacements, because of original variation of beds, and because of difficulty of confining sampling only to the recrystallized zone; but in some cases there is a remarkable tendency toward the constancy of silica-alumina ratios in comparison of original limestones and the supposedly recrystallized phases. The ratio is not absolutely maintained but the variation in the silica-alumina ratio is slight as compared to the variations which are found in the parts of the contact zone in which materials have been clearly introduced. It would be remarkable if substances brought in at random from magmatic sources should approximate so closely the composition of residual impurities of limestone. A most striking case of this, which has been worked out quantitatively on a large scale, is the contact metamorphism of cherty iron carbonates by great masses of gabbro in the Lake Superior country. Here the iron-silica ratio of the altered phase corresponds almost exactly with that of the original carbonate rock, the change being merely an elimination of carbon dioxide. Analyses have been taken from many thousands of samples brought up in drill cores and in continuous sections across the formation.[1]

(2) Secondary silicates of contact zones have often been found to be localized along cherty beds or around fragments of chert in the carbonate. Again the Lake Superior region furnishes an illustration in that cores brought from a depth of many hundred feet, where there has been no chance of surface alteration, and at some distance from the intrusive, show the development of secondary iron silicates, principally grünerite along contact of carbonate

[1] Van Hise, C. R., and Leith, C. K., The geology of the Lake Superior region: Mon. 52, U. S. Geol. Survey, 1911, pp. 204 and 546.

and siliceous layers, in a rock which is so dense there is little or no possibility for the introduction of these substances from without. The ratio of silica to iron has been almost exactly maintained. And yet these are clearly developed under influence of intrusives.

(3) The similarity of secondary silicates in limestones and marbles far removed from igneous contacts to some of those developed at contacts is also suggestive evidence of recrystallization along contacts.

(4) Elimination of carbon dioxide and lime is postulated under either hypothesis, "replacement" or "recrystallization." The natural consequence of elimination is recrystallization of the residual materials, whether or not these are supplemented by accessions from magmatic sources. There is no good *a priori* reason why accessions should always exactly balance elimination, especially when the physical conditions of intrusion are considered—and there is no satisfactory quantitative proof that they have. Under physical conditions which have been supposed to attend the earlier stages of intrusion of a magma it is easy to conceive of pore spaces caused by elimination to be closed as fast as formed, thereby reducing volume, and, in fact, it is usual to conceive of the pressure actually being a factor in the elimination.

Barrell [1] concluded from study of contact metamorphism of the Maryville batholith that

> "the causes preventing the addition of solid material are doubtless to be found in the fact that the gases must escape and that while this is taking place recrystallization is going forward; so that by the time relief is obtained from internal pressures and heated waters would be able to make their way inward, the rock is already dense and crystalline."

Under the replacement hypothesis we find it necessary to assume that whatever the pressure conditions were, whether tending to close up openings or not, the materials taken out and those introduced were delicately balanced in volume; that just enough is introduced in any one place to take the place of that which is left.

(5) The reduction in volume required by the recrystallization hypothesis cannot in many cases be disproved. So far as original

[1] Barrell, Joseph, The physical effects of contact metamorphism: Am. Jour. Sci., vol. 13, 1902, p. 292.

textures are retained, as they are in some districts, then it is possible to infer that the volume has not been considerably reduced, and, therefore, that elimination has not taken place except by equivalent introduction of new materials. But the supposedly recrystallized substances are often in a structurally amorphous zone which may well be the residual of an original mass many times greater. Opponents of the recrystallization hypothesis have argued that the necessary elimination of substances, and consequent reduction of volume, is too large to be reasonable. The reasonableness or unreasonableness of this is a difficult point to argue in the absence of specific proof. Elimination is equally necessary to the alternative hypothesis of introduction of the materials from magmatic sources. Without elimination it is necessary to assume an enormous increase in volume to take care of enough new material to give an average composition of the contact phase.

Barrell [1] has argued that the reduction in volume in the Maryville district is limited by the expulsion of only enough carbonic acid to free lime to combine with the siliceous impurities. In this case it might reach 50 per cent. Leith and Harder,[2] in their study of the contact deposit of the Iron Springs district of Utah, saw no reason to impose this limitation and argued the possibility of even greater reduction in volume.

(6) Discrimination of two phases of contact metamorphism is essential to an interpretation of conditions of formation of contact zones.

Goldschmidt,[3] in his classic work on contact metamorphism in the Kristiania region, distinguishes two phases, an inner zone of "normal contact metamorphism" and an outer zone of "pneumatolytic contact metamorphism." He believes the inner zone to be formed before the magma began to solidify without any notable fusion of the invaded rocks and without addition of water from the magma. He estimates the temperature to have been

[1] Op. cit., p. 285.

[2] Leith, C. K., and Harder, E. C., The iron ores of the Iron Springs district, southern Utah: Bull. 338, U. S. Geol. Survey, 1908.

[3] Goldschmidt, V. M., Die Kontactmetamorphose im Kristianiagebiet. (Videnskapsselskapets Skrifter, I. Mat.-Naturv. Klasse, 1911, No. 1.) Kristania, 1911. Review by F. L. Ransome, Econ. Geol., vol. 6, 1911, pp. 707–708.

between 1,000° and 1,200° C. and the depth about 5,000 feet, or a pressure of 400 atmospheres. The outer zone is formed later, probably during the crystallization of magma or even later, as some parts of the igneous mass were solid and were themselves effected by this kind of metamorphism. In this later stage comes the introduction of the metals, alkalies, chlorine, fluorine, sulphur, and silica. Iron-rich silicate rocks are produced by the addition of iron and silica to the limestones.

Students of contact metamorphism may to much advantage study the mathematical theory of heat conduction as applied to an igneous contact. We are indebted to Professors Ingersoll and Zobel [1] for a discussion of the principles of heat flow from an igneous rock of given dimensions into surrounding limestones. Their conclusions, which seem to be well based on general physical principles, are especially interesting in showing the remarkably slow progress of a heat wave into the limestone. Quoting from Ingersoll's and Zobel's discussion of a hypothetical case: [2]

> "The conclusions to be drawn from the curves are: first, that the cooling is a very slow process, occupying tens of thousands of years; second, that the boundary surface temperature quickly falls to half the initial value and then cools only slowly, and also that for a hundred or more years there is a large temperature gradient over only a few meters and a very slow progress of the heat wave; third, the maximum temperature in the limestone, or the crest (so to speak) of the heat wave, travels outward only a few centimeters a year. The mass behind it will then suffer a contraction as soon as it begins to cool, and the cracking and introduction of mineral-bearing material is doubtless a consequence of this."

Especially significant is the inference from the curves of heat flow that in advance of the heat wave the rock is tending to expand, therefore, to be compressed, whereas, following it during a long period of time there is contraction and the development of cracks. These conditions seem to favor two principal phases of contact metamorphism.

As the igneous mass advances into limestone it presumably is

[1] Ingersoll, L. R., and Zobel, O. J., An introduction to mathematical theory of heat conduction with engineering and geological applications, 1913, Ginn & Co.

[2] Loc. cit., pp. 128-129.

exerting mechanical pressure, judging by deformation at some contacts, and at the same time sending out heat into the surrounding rock, which, itself, increases the pressure. It is difficult to avoid the conclusion that for a time at least the adjacent rocks are under considerable pressure and that this pressure would favor elimination. It does not seem at all necessary or probable that under pressure this elimination should be immediately followed by introduction of other substances from the magma, or putting it in another way, that substances from the magma should always so closely follow elimination as to replace molecule by molecule the original materials and thereby prevent any reduction of volumes. As the crest of the heat wave advances into the surrounding limestone, lower temperatures follow, with the result that there is contraction and the development of openings. This contraction may affect not only the limestone but the intrusive itself. Into such openings the magmatic solutions may freely enter, and there are deposited the ores and some of their gangue materials. At the same time these solutions may replace the materials of the surrounding rock to a greater or less extent.

That contact metamorphism of limestone has been accomplished in successive phases has been pretty well proved at certain contacts. It seems probable that when attention is directed specifically to this feature it may be found at others. The first phase seems to be characterized by the production of an amorphous, homogeneous, silicate mass, not definitely associated with fissures. In some cases this is discriminated sharply from, in other cases it merges gradually into, a phase characterized by sulphides and other ore-bearing minerals with their gangue materials, which occur much more largely in fissures. These fissures may often be seen to traverse the silicate zone of the first phase. The minerals of the later phase, both because of their composition and because of evidence of their transportation, cannot be regarded as recrystallizations of materials in place. They afford evidences of introduction from magmatic sources.

The two phases of alteration may merge one into the other both in time and place. The later phase may be expected to obliterate to some extent the earlier phase. Ordinarily the later minerals differ from the earlier ones, but certain silicates, quartz, and other minerals, may be common to both.

INFLUENCE OF PRESSURE ON ESCAPE OF CARBON DIOXIDE IN CONTACT METAMORPHISM OF CARBONATES

Whether contact alteration of limestone is by elimination of lime and carbon dioxide alone or by elimination combined with replacement of substances from the igneous mass, avenues of escape must be open for the carbon dioxide or calcium or both. Calcium carbonate is ordinarily decomposed at 500° C. but high pressure will prevent it from decomposing at much higher temperatures. Inferentially, then, the possibilities of escape of carbon dioxide and lime should have an effect on the nature and extent of the alteration. If at great depth or under impervious cover, it might be inferred that alteration to silicates, however it is accomplished, would go on more slowly than at places where the carbon dioxide and lime could escape freely. This may be the reason for the existence of coarsely crystalline slightly silicated marbles at some points of igneous contacts, although these have also been explained as due to the absence of mineralizing substances from the magma at some points, or as the result of anamorphism of lime carbonates which were unusually pure. Barrell [1] concluded, from study of contact effects of the Maryville batholith, that

> "carbonic acid is only expelled where the siliceous impurities of the limestone are sufficient to combine with the lime set free, forming lime silicates. This ability of deeply buried limestones to retain their carbonic acid when intensely heated, if free from other impurities, has been noted by a number of observers, and is well shown in the Elkhorn district, Montana, where marbles in some cases exist within two to four feet of the granite contact, having retained their carbonic acid under temperatures which at the surface would have led to its immediate expulsion."

The precise influence of this factor of ease of escape of substances at the contacts upon the alterations of limestone has not yet been ascertained from field study. Results of significance are to be looked for when this study is made.

A SPECIAL CASE OF CONTACT METAMORPHISM OF LIMESTONE

A special case of contact metamorphism of limestone on an enormous scale needs further mention. At contacts of certain of

[1] Barrell, Joseph, The physical effects of contact metamorphism: Am. Jour. Sci., vol. 13, 1902, p. 285.

the Grenville limestones with Laurentian granites, in Ontario and Quebec, the limestone has been described as changed to amphibolites through many hundreds of feet.[1] Certain large areas of amphibolite are clearly the result of dynamic metamorphism of certain basic igneous rocks, and the question has been raised as to what extent the amphibolites represent altered limestones and to what extent they may represent the altered igneous rocks. The fact remains, nevertheless, that amphibolites have been

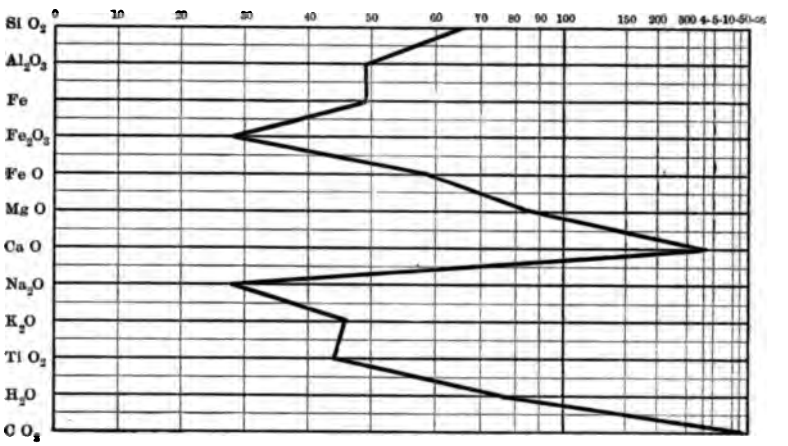

FIG. 13. Representing losses and gains of constituents in contact metamorphism of Grenville limestones. Analyses by Adams and Barlow.

proved on a large scale to be the result of contact metamorphism of limestone by granites. Comparison of analyses (Fig. 13) has suggested to Adams and Barlow that the amphibolites are developed by the direct replacement of the limestone by substances in the hot solutions from the granite. Certain it is that these amphibolites were developed outside of the granitic mass under differential pressures, for they show parallel arrangement of constituents which is maintained without modification even in blocks which are enclosed in the granitic mass. This fact prevents the conclusion that the amphibolite rocks in the granite along the

[1] Adams, F. D., and Barlow, A. E., Geology of the Haliburton and Bancroft areas, Ontario: Mem 6, Geol. Survey of Can., 1910; Adams, F. D., 'On the origin of the amphibolites of the Laurentian area of Canada: Jour. Geology, vol. 17, 1909, pp. 1-18.

contact are the result of replacement of limestone blocks after these had been engulfed in the granitic magma. If the amphibolites develop principally in originally impure parts of the limestone, as held by Adams and Barlow, then this fact suggests development of silicate from the substances in the limestone rather than from substances contributed from without. Inferences from analyses of the limestones and amphibolites in Fig. 13 seem to the writers to be as well in accord with one hypothesis as another. More adequate sampling and analysis are needed.

CONCLUSION AS TO CONTACT AND THERMAL METAMORPHISM OF LIMESTONE

The problems of thermal and contact metamorphism of limestone have been sufficiently well defined to make their solution seem easy. In fact, a comparatively small amount of quantitative data would answer some of these questions. It is most surprising on going through the literature to find how little there is of quantitative data. There have been published up to the present time only fifteen sets of analyses showing definitely what has happened in the change from the fresh to the altered rock. Some of these analyses are incomplete and some of the specimens were collected under conditions which throw doubt on the results. The contact phases must be studied in bulk, rather than in hand specimens, before some of these questions can be answered. The geologist is in the habit of bringing home specimens and having them analyzed later, using such analyses as typical of the whole. It would be more desirable for him to follow the method of mining engineers in sampling ores and take for each analysis systematic samples over a large area, representing a general average. Also insufficient attention has been given to observational evidences of volume change, knowledge of which is essential to adequate interpretation of contact phenomena.

We may conclude that the silication of a limestone under anamorphism is accomplished, so far as present evidence goes, by (1) elimination of carbon dioxide, and to some extent lime, and by recrystallization of the remaining lime and siliceous impurities into silicates, and (2) by the addition of substances from outside; that quantitative evidence is yet quite insufficient to determine the relative importance of elimination of substances and the

addition of substances from outside; that so far as the quantitative evidence goes it shows clearly that both occur; that if the process is mainly by elimination, then the anamorphism means great reduction of volume—if by substitution, the volume change may be small, and that the chemical changes have been those of combination, requiring on the whole the addition of heat.

DOLOMITIZATION OF LIMESTONE

The dolomitization of limestone is described under a separate heading for the reason that it may be only partly an anamorphic change. Dolomites may result from original precipitation as chemical sediments; replacement of lime by magnesia may go on in contact with sea water; groundwaters may replace lime by magnesia or enrich the rock in magnesia by abstracting lime from crystalline carbonates.

Some evidence has been found for each of these methods of development of dolomite,[1] but there is wide range of opinion as to their relative importance.

(1) Perhaps the least decisive evidence is for the hypothesis that dolomite may be precipitated as a chemical sediment. Direct evidence for this is in the nature of the case difficult to secure. Dolomite may be experimentally reproduced under conditions that suggest this possibility. That it is not the common case is perhaps best shown by the irregularity of occurrence of dolomite in limestone beds, indicating replacement subsequent to deposition. It is found as the matrix or paste surrounding calcareous shells, indicating that it is at least later than these forms. Progressive increase in magnesium with depth in certain coral reefs is further evidence of change subsequent to deposition. In fact, all of the evidences for the two replacement hypotheses, discussed below, can be used negatively in consideration of the hypothesis of direct precipitation as a chemical sediment.

(2) That dolomite results from replacement of limestone by action of sea waters is indicated by general evidences of replacement of limestones and by the difficulty of explaining this replacement by the agency of meteoric waters at the surface. From

[1] See summary by Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, pp. 534–545.

borings in the coral atoll of Funafuti, J. W. Judd showed that there is a progressive increase of magnesia in depth.[1] The lower portions of the core, which were the earliest deposited and therefore have been acted upon the longest time, are the most completely changed. The abundant occurrence of dolomite in the cements between marine fossils, which have not been altered, indicates introduction of the dolomite after the accumulation of the fossil shells. That many of the replacements of limestone by dolomites are not caused by meteoric waters percolating from the surface, but were accomplished at an earlier stage seems to be indicated, according to Steidtmann,[2] by the following facts: Dolomite occurs irregularly in limestone beds not otherwise altered; dolomite does not occur along fissures or bedding planes or other openings notwithstanding the fact that there is clear evidence that the rock is permeable to meteoric waters; it is confined often to fairly definite horizons, not related to impervious layers or other conditions which would localize flow of waters, while the horizons are equally accessible to meteoric waters; it is uniformly associated with ferrous iron, and this uniformity of ferrous iron content would not be normal to meteoric waters; finally dolomites as a whole do not show the porosity that they should show as the result of replacement of crystalline limestone rocks. Worm borings in dolomites are very common, but otherwise the porosity does not seem to be higher than that of limestone. Theoretically the replacement should result in a reduction in volume of 12 to 14 per cent. The increase in density is from 2.72 to 2.85. Steidtmann argues that this absence of porosity may be explained on the hypothesis that the replacement goes on by addition of magnesium carbonate in solution in sea water to solid lime carbonate. There is, however, no experimental evidence on which any of these interpretations could be based. On the whole, the evidence for this hypothesis of origin of dolomites seems to be the best substantiated of any of the hypotheses, although it is to be admitted that it is based largely on negative evidence from the difficulty of explaining dolomitization on the other hypotheses.

[1] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, p. 542.

[2] Steidtmann, Edward, The evolution of limestone and dolomite: Jour. Geol., vol. 19, 1911, pp. 334–336 and personal communication.

(3) There is some evidence that dolomites are formed by direct replacement of limestone by ground waters. The occurrence of dolomite veins, as in the lead and zinc fields of Missouri, is evidence of this kind.

Daly [1] has compiled tables to show that dolomites are in larger proportion compared to limestones in the older rocks than in the younger rocks. This fact has been cited in support of the hypothesis that dolomitization is a secondary process brought about by meteoric waters, it being assumed that older rocks, on the whole, have had more time to undergo these metamorphic changes. Daly,[2] however, infers that lime has been less abundant in the early oceans than it is now.

An interesting fact brought out by Steidtmann [3] from his observations on a wide variety of dolomitic limestones, with the aid of stains, is that there seems to be no gradation between limestone and dolomite, but usually an intermixture of clean dolomite in limestone.

ENDOMORPHIC CONTACT EFFECT AT LIMESTONE CONTACTS

At a considerable number of contacts practically no mineralogic change can be observed in igneous rocks near limestone, except possibly a change in grain. In some cases there has been described the development of garnet, diopside, amphibole, and other silicates near the contact, which has been ascribed to absorption of some of the limestone. There is but little quantitative evidence of any change in composition which may have taken place. Whether types of igneous rocks representing any considerable mass have resulted in the absorption of limestone remains to be proved. The abundant amphibolites along the contact between granite and limestone in the Grenville series have been ascribed to the interaction of the two series, and presumably in one sense this might be regarded as an endomorphic effect, but this is complicated by the question as to how far the amphibolites may be the result merely of recrystallization of impurities in the limestone,

[1] Daly, R. A., First calcareous fossils and the evolution of the limestones: Bull. Geol. Soc. Am., vol. 20, 1909, p. 165.

[2] Daly, R. A., The limeless ocean of pre-Cambrian time: Am. Jour. Sci. vol. 23, 4th ser., 1907, p. 113.

[3] Personal communication.

as they are undoubtedly in certain beds, and to what extent these amphibolites are the result of the metamorphism of certain basic igneous rocks, which has been proved in some cases. It has been suggested that certain of the rare high alkaline rocks, like nepheline syenites, may be the result of interaction between granite and limestone, the theory being that the fused lime has taken out much of the silica and sunk, leaving high alkaline parts of the magma deficient in silica to crystallize as nepheline. Daly [1] has summarized some 95 occurrences of nepheline syenite, at or near limestone contacts, although he recognizes the existence of many nepheline syenites at places where no limestone has been found. This interesting hypothesis needs further tests before acceptance. It is a question whether the association with limestones is any more prevalent than the association with other types of rocks.[2]

[1] Daly, R. A., Origin of the alkaline rocks: Bull. Geol. Soc. Am., vol. 21, 1910, pp. 92–109.

[2] Cross, Whitman, Lavas of Hawaii and their relations: Prof. Paper 88, U. S. Geol. Survey, 1915, p. 90.

CHAPTER IV

DYNAMIC AND CONTACT METAMORPHISM OF IGNEOUS ROCKS

DYNAMIC METAMORPHISM OF IGNEOUS ROCKS ("ROCK FLOWAGE")

Gneissic structure may be developed in the igneous rocks as an original or primary structure forming "protogene gneisses," and, by minute injection of igneous rock along parallel surfaces, forming the "injection gneisses." Igneous rocks may also be rendered schistose or gneissic by dynamic metamorphism. This section relates mainly to the last named group of schists and genisses. Attention is directed to the question of the katamorphic or anamorphic nature of the changes, and their relation to the metamorphic cycle. A few typical alterations are sketched for illustrative purposes.

A basic igneous rock becomes a chloritic, hornblendic or micaceous schist. The hornblendic and biotitic schists are especially characteristic of igneous contacts, and chloritic schists are characteristic of dynamic metamorphism. These schists differ in no essential respect from chloritic, micaceous and hornblendic schists derived from the anamorphism of certain sediments. The conspicuous mineral changes developed in the basalt are augite to hornblende, augite to chlorite, augite to hornblende and chlorite, hornblende to hornblende and chlorite, hornblende to chlorite, in each case with the separation of lime and some silica. Basic feldspars alter into more acid feldspars, particularly albite, with a separation out of muscovite and lime silicates of the epidote and zoisite type, and ultimately of lime carbonate. Where the albite and epidote-zoisite type of alteration products are closely interwoven, the combination is called *saussurite*. Where the hornblende is in more or less fibrous form with peripheral or radial arrangement relative to the original augite, it is called

uralite. Olivine changes to tremolite, anthophyllite and talc. On considering the chemistry of these changes, it appears that they are all in the direction of breaking down of complex molecules. Calcium salts separate as by-products of almost every change, as for instance from augite to hornblende, hornblende to chlorite, from basic feldspar to acid feldspar and its by-products. These calcium salts may remain as silicates or as carbonates or may be carried off. Water is introduced, for the amphiboles, chlorites, and micas are hydrous as compared with the augites and feldspars. Chemical analyses illustrating this change are rare. Those published by Williams [1] seem to be of the best selected, but even these rocks are more or less weathered. All available pairs of analyses are summarized on Plate XIII. Such as they are, they seem to indicate a loss of calcium, a gain of water, and perhaps a slight gain of soda and potassium.

Following is a list of references to the analyses platted:

1. } Gabbro, Michigan. Williams, G. H., Bull. 62, U. S. Geol. Survey,
1a } p. 76.
2. Greenstone, Michigan. Williams, G. H., Bull. 62, U. S. Geol. Survey, p. 91.
3. Gabbro diorite, Michigan. Williams, G. H., Bull. 62, U. S. Geol. Survey, p. 89.
4. Granitoid gneiss, southwestern Minnesota, Hall, C. W., Bull. 157, U. S. Geol. Survey, p. 68.
5. Greenstone, Marquette district, Michigan. Leith, C. K., unpublished analyses.

Dynamic alteration of acid igneous rocks usually produces hornblendic, micaceous, chloritic schists. The principal mineral change, in addition to those described in preceding paragraphs, is the conversion of the acid feldspar into white mica, with liberation of quartz. Tendencies to produce mica, hornblende, and chlorite, at the expense of other minerals, or with the elimination of bases and quartz, are noted throughout. The essential textural change is the parallel arrangement of mineral constituents, segregation of like minerals into bands, and a tendency to coarsening of grain, although at early stages the grain may become finer, due to granulation of primary minerals. (See pp. 174–175.)

[1] Williams, G. H., The greenstone schist areas of the Menominee and Marquette regions of Michigan: Bull. 62, U. S. Geol. Survey, 1890.

The continuation of the above described processes may yield a coarsely banded rock with abundant secondary feldspar, which is ordinarily termed *gneiss*. There is no sharp definition between a gneiss and a schist in regard to minerals, textures, or conditions under which they are produced. Rocks on the border line may be called schist by one observer and gneiss by another. While gneiss is often seen to represent a stage of metamorphism beyond that of a schist, it does not follow that all gneisses have gone through the schist stage. Some may develop directly from igneous rocks, especially by granulation, as described below.

The origin of schists and gneisses is discussed from another point of view in Chapter I, Part III.

Does the Dynamic Metamorphism of Igneous Rocks Represent Anamorphism or Katamorphism?

It will be noted that the products formed during the crushing of igneous rocks are essentially the same as those which are developed by the anamorphism of sediments. Yet, when they come from igneous rocks they represent relative breaking down of molecular structure, whereas from the sedimentary rocks they represent a relative building up. Is the change, then, katamorphic for the igneous rocks and anamorphic for the sediments? The answer to this question may perhaps be indicated by the significant fact, which has been noted in many fields, that schists develop most readily from the weaker or incompetent parts of igneous masses, which are likely to be weathered or thermally altered. For instance, the finer-grained [illegible] and surface phases of basalts pass into schists more commonly than the massive phases. The more easily destroyed parentage of the ellipsoids, so common in basalt flows, was noted under dynamic metamorphism. Many basalts are [illegible] and chloritized before the schistose structure is developed. This is notably the case in the Keweenaw [illegible] Lake superior region and Canada, where [illegible] depths of many hundreds of feet [illegible] which certainly went [illegible] solutions connected with [illegible] Surficial zones in rhyolites [illegible] while the fresh zone [illegible]

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The origin of schists and gneisses is discussed from another point of view in Chapter I, Part III.

Does the Dynamic Metamorphism of Igneous Rocks Represent Anamorphism or Katamorphism?

It will be noted that the products formed during the mashing of igneous rocks are essentially the same as those which are developed by the anamorphism of sediments. Yet, when they come from igneous rocks they represent relative breaking down of molecular structure, whereas from the sedimentary rocks they represent a relative building up. Is the change, then, katamorphic for the igneous rocks and anamorphic for the sediments? The answer to this question may perhaps be indicated by the significant fact, which has been noted in many fields, that schists develop most readily from the weaker or incompetent parts of igneous masses, which are likely to be weathered or thermally altered. For instance, the finer-grained, more tuffaceous and surface phases of basalts pass into a schist more commonly than the massive phases. The more easily altered peripheries of the ellipsoids, so common in basalt, first become schistose under dynamic metamorphism. Many basalts are pretty well carbonated and chloritized before the schistose structure is developed. This is notably the case in the Keewatin basalts of the Lake Superior region and Canada, which are carbonated and chloritized at depths of many hundreds of feet from the surface by alterations which certainly went on before their burial, perhaps by the hot solutions connected with the extrusion of the basalts themselves. Surficial zones in rhyolite flows are likely first to become schistose, while the fresh zones remain comparatively unaltered. The oc-

currence of schists along shear planes of fracture has been frequently noted. In these places katamorphic alterations of igneous rocks along the fissures may have occurred before the schistosity was developed. Surface volcanics in general may be more likely to be katamorphosed prior to the development of schistose structure than are plutonic rocks, but there are many exceptions.

It would therefore seem that at least some of the schists developing from an igneous rock have really not come directly from the fresh igneous rock but from its katamorphosed phases. One of the causes of the easy development of schists from katamorphosed igneous rocks is that these rocks are on the whole weaker and tend to yield to rock flowage more easily than fresh and massive igneous rocks, the weakness being due not only to the katamorphic products, but to the fissures and textures which gave opportunity for weathering. So far as schists and gneisses result from katamorphosed igneous rocks, the process differs in no essential respect from the anamorphism of sediments, which represent the end-products of weathering of igneous rocks. In other words, the metamorphic cycle, from original igneous rock through katamorphic products to schists, is just as certainly represented in these cases as development of schists from completely weathered igneous rock which has been distributed as sediments. This hypothesis explains the common occurrence of schists along fissures and along zones of earlier weathering. It explains the fact that the change is one of simplification, separation out of lime, and hydration, so far as comparison of schist and igneous rock is concerned. Anamorphism tends to bring the composition back toward that of the igneous rock, but so far as it falls short of reaching this composition the differences are the same in kind as those between igneous rocks and katamorphic end-products.

If schists and gneisses develop commonly from katamorphosed phases of igneous rocks, it does not follow that all igneous rocks pass through an intermediate katamorphosed state to become schists and gneisses. It is entirely conceivable that the schist may develop from fresh igneous rock which has not been previously katamorphosed, but there must be some way for simultaneous introduction of katamorphic agents, especially water and carbon dioxide, with partial elimination of certain constitu-

ents, for hydrous silicates are developed and carbonates and quartz freed by the destruction and simplification of primary silicates, which implies that the change must take place under conditions allowing access of katamorphic agencies. When the schist is developed along a shear zone, the katamorphism and development of schistose structure may be almost simultaneous, or at least closely successive, processes.

Where water and carbon dioxide cannot get in and lime cannot get out, is it still possible to produce a schist? In such a case, it would seem that a rock in yielding to differential pressure probably would undergo other mineral changes which do not require these substances. Granulation or recrystallization of primary minerals may then occur. The anorthosite at Montreal described by Adams [1] has apparently undergone a mashing, developing a banded or gneissic rock of a lighter color, in which original feldspars and augites are granulated and stream out into bands without any conspicuous change in mineral composition or the development of hydrous minerals. Plutonic rocks which have never reached the surface might well be altered under such anhydrous conditions. Many secondary gneisses are distinguished from schists by the lack of development of hydrous minerals, lack of evidence of mineral change, and frequently by conspicuous evidence of granulation. They are crystalline anhydrous rocks. Had there been any katamorphosed products or water or carbon dioxide present during their mashing, it is difficult to see why mica or chlorite or hornblende should not have developed. The absence of these minerals is taken to imply the absence of katamorphic alterations or of water or carbon dioxide.

It is concluded, in summary, that inasmuch as the mica, chlorite and hornblende of the schists formed from igneous rocks are hydrous minerals and imply in their development the agency of water and carbon dioxide for the elimination of lime, the schist was developed under conditions which permitted introduction of water and carbon dioxide and elimination of lime, or from rocks in which these changes by water and carbon dioxide had been made before the schist was developed. Field evidence

[1] Adams, F. D., Report on the geology of a portion of the Laurentian area lying to the north of the island of Montreal: Ann. Rept. Geol. Survey of Can., vol. 8, pt. J, 1896, pp. 85 et seq.

shows in some cases that schist clearly develops from previously katamorphosed portions of igneous rocks. Schists thus developed therefore are the results of anamorphism of katamorphic products, just as are schists derived from sediments. Products of these processes in the extreme may be gneisses with secondary hornblende or biotite. Some granular gneisses may be developed by the mashing of fresh igneous rocks under conditions which do not allow the introduction of carbon dioxide and water necessary to make the hydrous silicates so characteristic of schists. In such gneisses the original minerals figure more largely, granulation of original minerals is more conspicuous, development of hydrous minerals implying the elimination of lime is relatively less conspicuous. Gneisses are produced also by the continuation of anamorphism of the sediments beyond the schistose stage or distortion of original textures. In general the development of schists and gneisses, then, from igneous rocks, represents a net katamorphic simplification of molecules or mechanical breaking down of texture, not because katamorphism alone has been operative, but because katamorphism has accomplished results prior to or during the change, which anamorphism has not completely eradicated.

CONTACT AND THERMAL ALTERATIONS OF IGNEOUS ROCKS

The development of schists and gneisses from igneous rocks goes on both under dynamic metamorphism and along deep-seated igneous contacts. Present information does not allow us to discriminate very closely the products of these two environments. Near plutonic masses the changes are likely to have gone to an extreme. Hornblende and biotite (the latter often in porphyritic forms) and secondary albite (also in porphyritic form) are likely to become conspicuous. Whether or not material has been introduced from the igneous rocks along the contacts, analyses do not show, although the abundant development of albite suggests here the introduction of soda, just as in sediments at contacts with igneous rocks. By fusion augite and olivine are developed from hornblende.

Along open fissures at considerable depth igneous rocks suffer alteration and replacement through the agency of hot mineral-bearing solutions. Quartz, pyroxenes, amphiboles, garnet, apa-

tite, ilmenite, magnetite, tourmaline, topaz, the brown and green micas, spinels, and the soda-lime feldspars, are formed in various proportions. Sulphides and arsenides are commonly associated with these phases and constitute important ore bodies. Some of these minerals contain boron, chlorine or florine, which are important constituents of high temperature solutions, of mineralizers, known to be agents in effecting alterations of this kind. The temperature has commonly been below 575° C., as determined by the crystallographic properties of the quartz, which indicate crystallization below its inversion point.

In Plates XIV and XV are expressed the changes in chemical composition in alterations of this class. Following is a list of references to the analyses platted on Plate XIV.

8. Quartz monzonite, Rimini, Montana. Knopf, Adolph, The tourmalinic silver-lead type of ore deposit: Econ. Geol., vol. 8, 1913, p. 115.
12. Monzonite porphyry, Clifton-Morenci district. Lindgren, W., The copper deposits of the Clifton-Morenci district, Arizona: Prof. Paper 43, U. S. Geol. Survey, 1905, p. 168.

20a, 20b, 20c. Granites and greisens developed from them. Lindgren, W., Mineral Deposits, p. 621.

21. Grainsgill granite and greisen, Grainsgill, England. Harker, A., The Grainsgill greisen: Quart. Jour. Geol. Soc. London, vol. 51, 1895, p. 141; Econ. Geol., vol. 7, 1912, p. 244.

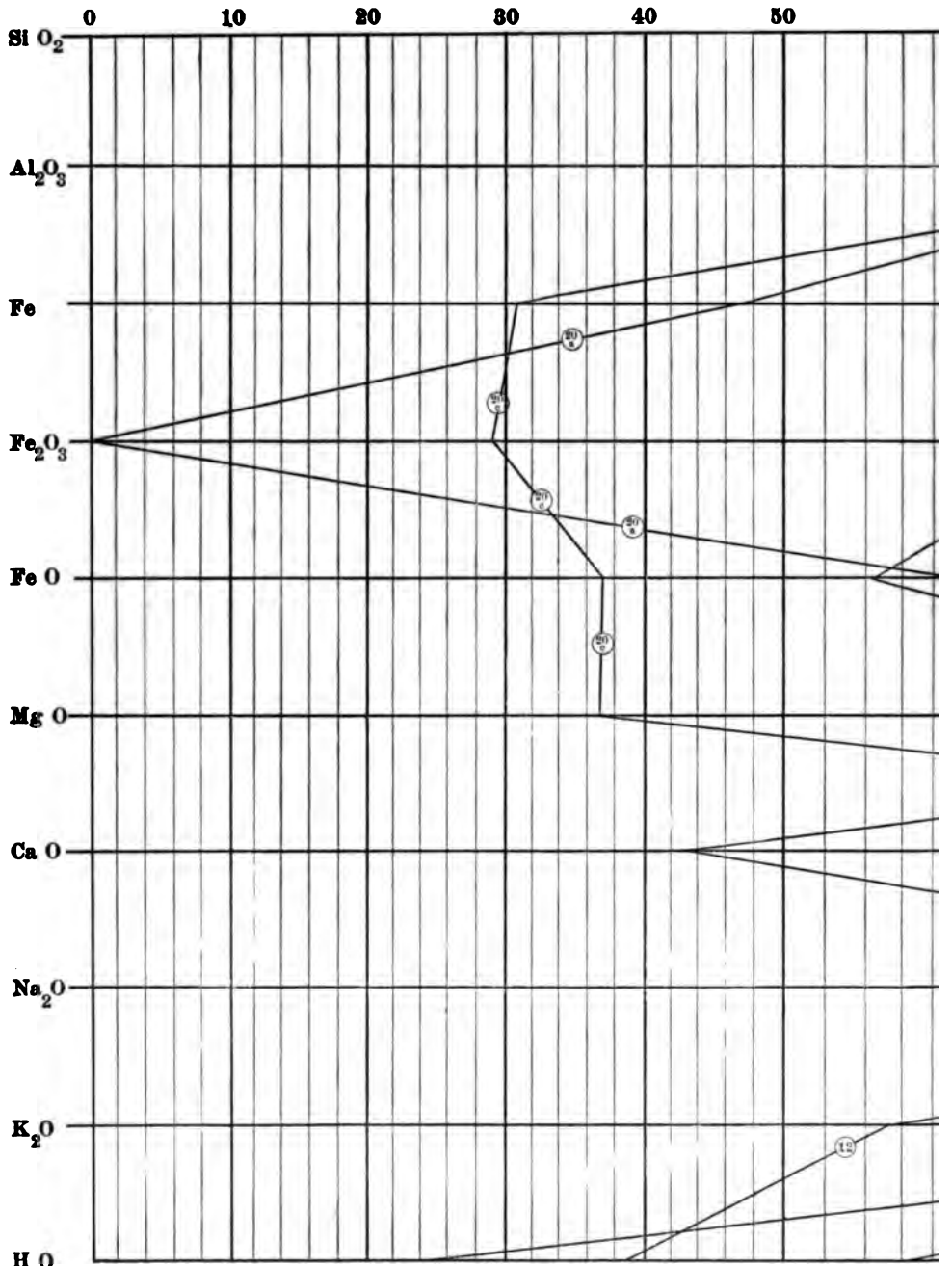
Following is a list of references to the analyses platted on Plate XV:

5. Amphibole schist, Kalgoorlie, West Australia. Lindgren, W., Metasomatic processes in the gold deposits of western Australia: Econ. Geol., vol. 1, 1906, pp. 536–537.
9. Amphibolite, Alaska. Knopf, Adolph, The Eagle River region, southeastern Alaska: Bull. 502, U. S. Geol. Survey, 1912, p. 39.
10. Diorite, Gold Creek, Alaska. Lindgren, W., Mineral Deposits, p. 646.

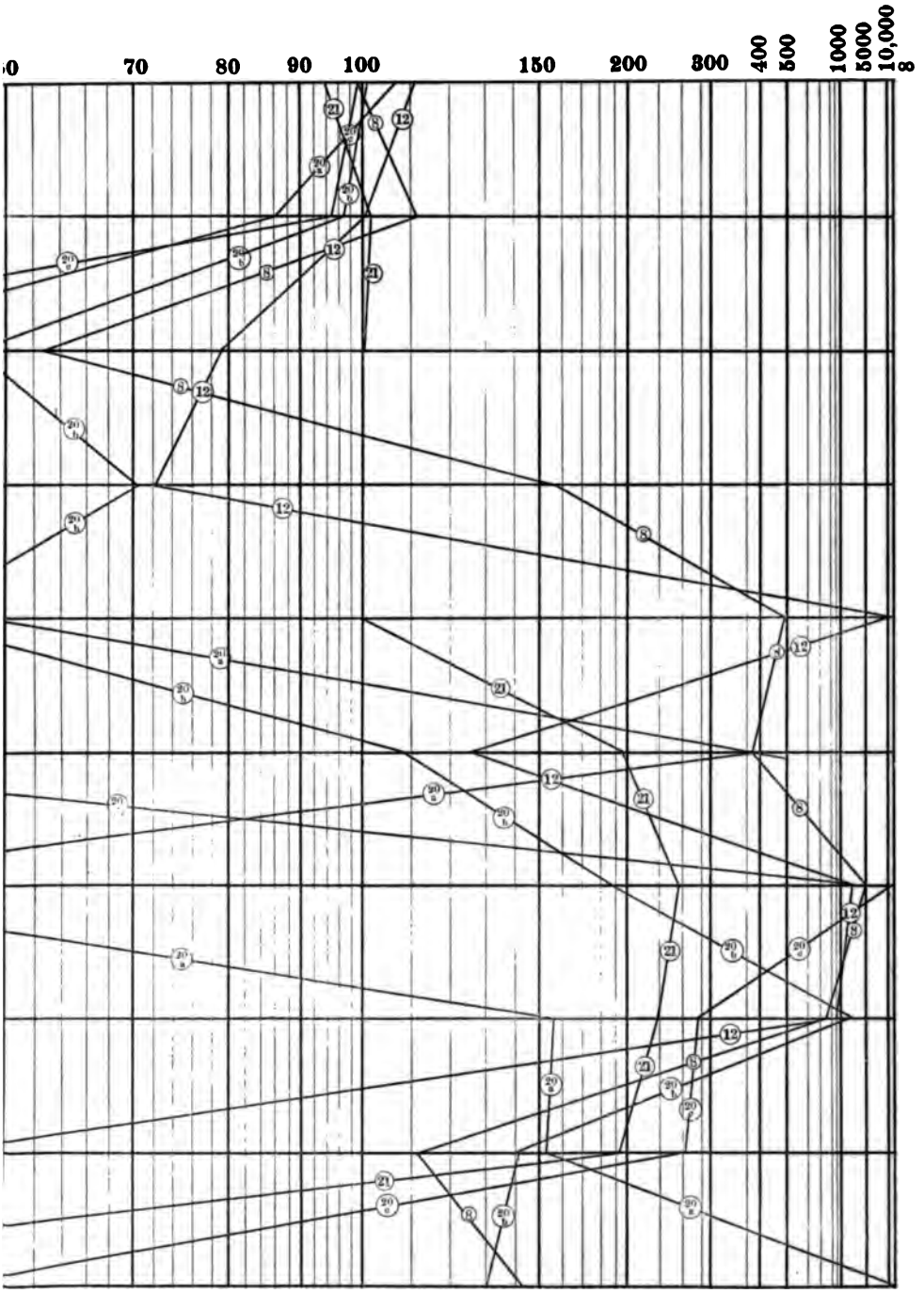
It will be noted that iron and ferric oxide are generally gained, while silica, magnesia, lime, soda, and water, are generally lost. Fluorine is not shown in this diagram but it is generally present. This alteration differs most distinctively from that at intermediate and shallow depths by the less conspicuous introduction of potash. (See Pls. V and VI.)

The alterations by hot waters in this deep zone are essentially anamorphic in their general tendency and relationships. Nearer the surface hot waters accomplish changes essentially katamorphic or in such association with katamorphic alterations that they may be more conveniently described in that connection.

PLATE XIV

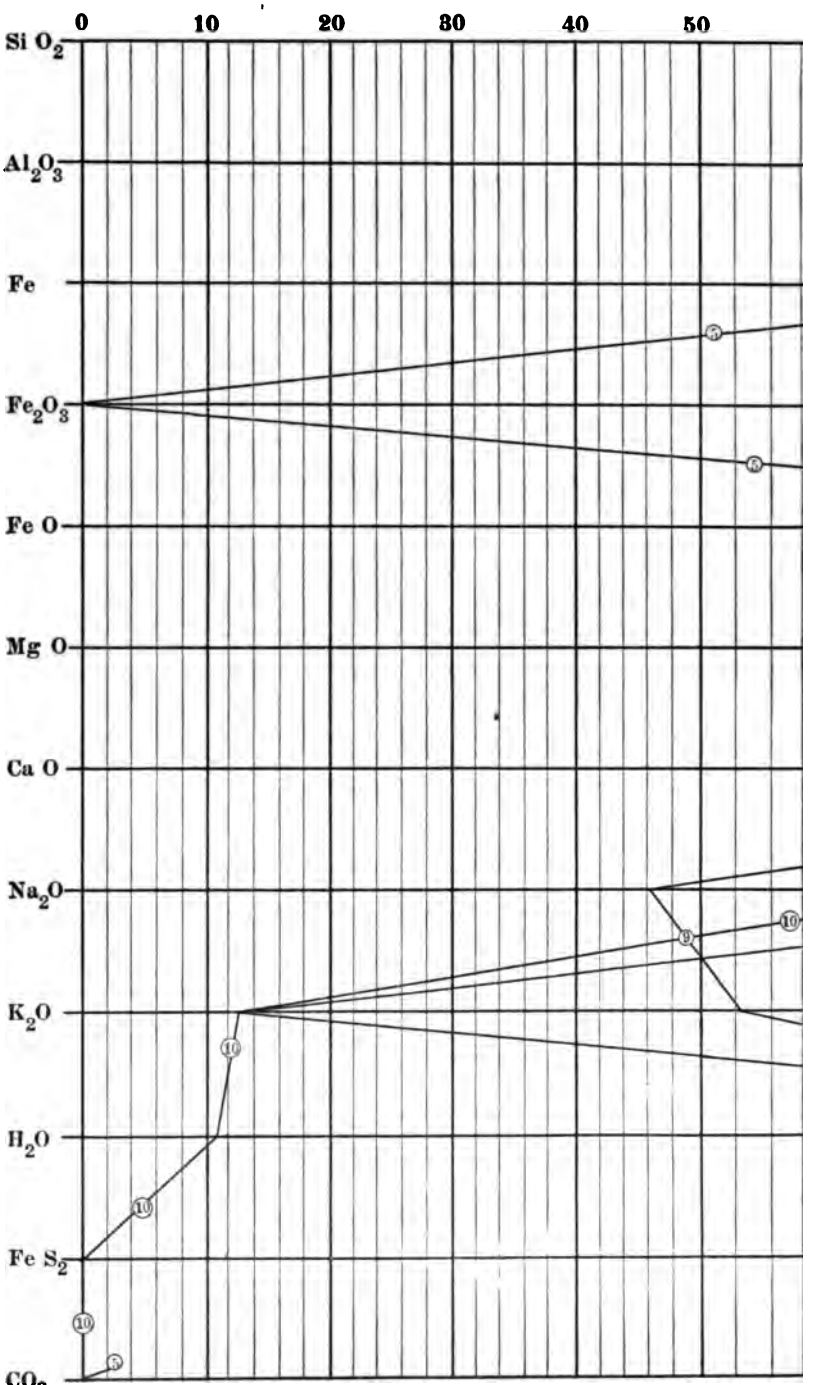


Representing losses and gains of constituents in the contact and thermal
of a pair of analyses, n

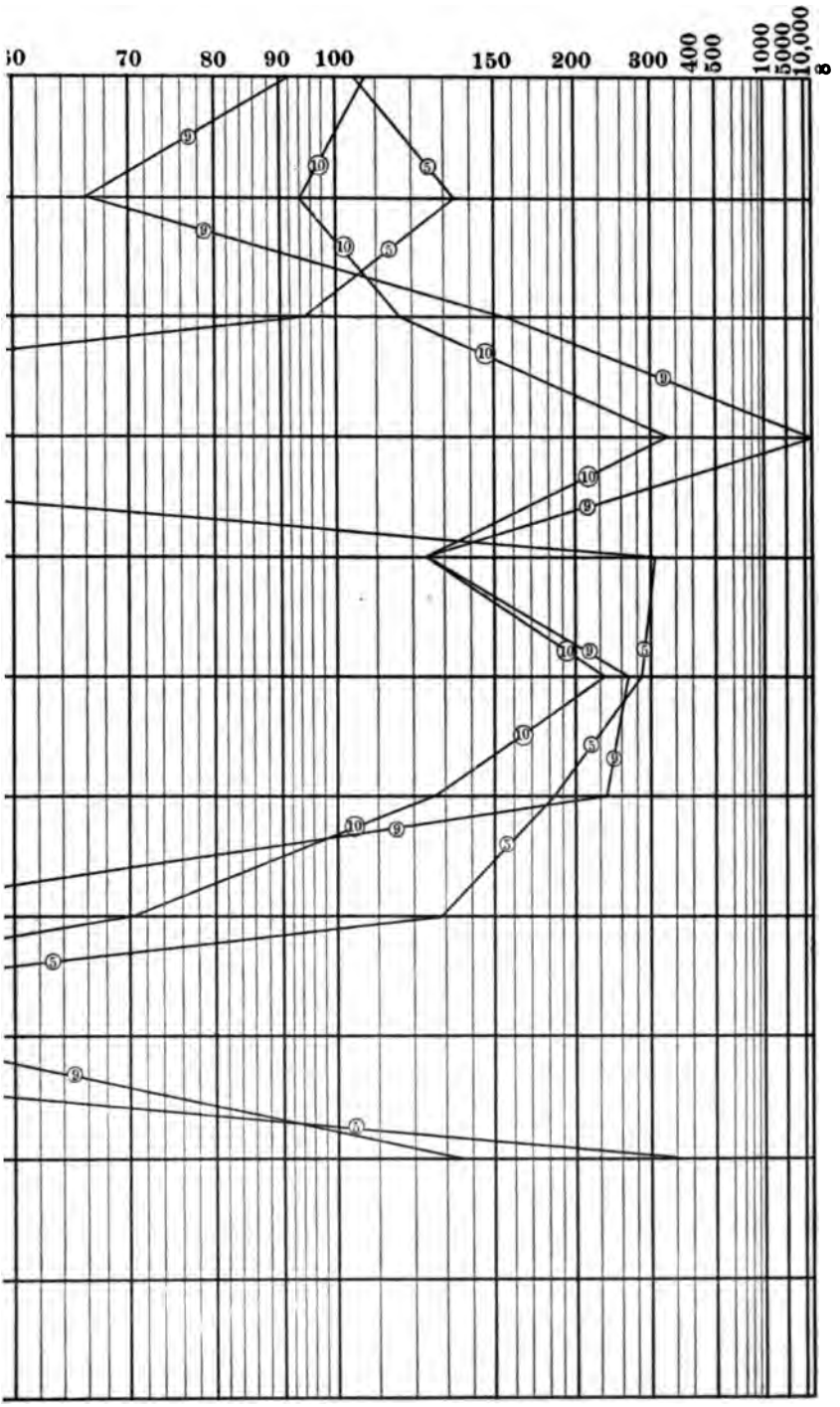


lteration of acid igneous rocks. Each broken line represents comparison
bered for identification.

PLATE XV



Representing losses and gains of constituents in the contact
represents comparison of a pair



d thermal alteration of basic igneous rocks. Each broken line
nalyses, numbered for identification.

CHAPTER V

ANAMORPHISM IN ITS RELATION TO MINERAL DEPOSITS

ANAMORPHISM AND METALLIC MINERAL PRODUCTS

The question of the relationship between ore deposits and anamorphic processes presents itself in several phases: (1) Cementation of ores, which includes some processes of secondary enrichment; (2) the anamorphism by rock flowage of ore bodies previously formed; (3) the segregation of ore bodies during anamorphism by rock flowage; (4) the development of ore bodies by thermal waters in connection with intrusives.

(1) Conspicuous illustrations of enrichments of ores by cementation are secondary sulphide enrichments in copper and zinc deposits. To a less extent evidences of enrichment by cementation appear in silver, iron, and other deposits. Secondary enrichment of this type is complementary to the leaching of ores in the belt of weathering and is largely confined to a zone near the top of the water table. In these respects secondary enrichment is more nearly related to katamorphism than to anamorphism, but so far as this process tends to cement and solidify the rocks, it has anamorphic tendencies, contrasting with the disintegrating tendencies of katamorphism.

(2) The anamorphism of ore bodies by rock flowage tends to recrystallize them, to develop the characteristic silicates of rock flowage, to reduce porosity and volume, to dehydrate, to render the ores hard and crystalline, and sometimes schistose. No data are yet available to show that this process has tended greatly to benefit the grade of the ore or to segregate the ore more effectively. The general tendency seems rather to be to tie up the ore with its gangue in crystalline combinations, which makes the ore more difficult to treat either by mechanical or metallurgical processes. Hematites and limonites formed under surface conditions may be

rendered hard, crystalline, and schistose, as illustrated by the specular hematites of the Lake Superior region and by certain hematites and magnetites of Sweden and Norway. The Lake Superior specular hematites were formed as soft hematite and limonite on or near the erosion surface and were then deeply buried and anamorphosed. Presumably there has been some slight enrichment by elimination of water. The associated quartz has been thoroughly recrystallized to form much larger grains. Some of the iron has been reduced to magnetite. Porosity has been almost entirely eliminated and the ore has taken on a schistose structure. Where the hematites and limonites are associated with clays, as in the brown ores, the result is a crystalline schist containing small lenses of hematite and more or less magnetite. The gangue in which the ore lies cannot be easily separated, as it could when in the katamorphic condition. Ordinarily such deposits are of no commercial value. The residual deposits of the Cuban type when anamorphosed become dehydrated to hematite and magnetite, thoroughly recrystallized, with the development of secondary silicates such as enstatite. The Cle Elum magnetite deposits of Washington are supposed to have been so formed. They are much less easily handled than the Cuban ores, partly because of their highly crystalline character and partly because they have been folded, requiring underground mining and perhaps mechanical concentration rather than surface recovery with a steam shovel.

Bauxite deposits, because of their high content of combined water, are easily modified by anamorphic dehydration. The common aluminum hydrate of residual bauxite deposits is $Al_2O_3.3H_2O$. By dehydration this is reduced gradationally to corundum. Many of the European deposits which have suffered compression contain the lower hydrates, are more or less schistose, and less easily refined than the unanamorphosed material. In general anamorphism has a deleterious effect on bauxite, rendering it more difficult to work both mechanically and chemically, although relatively increasing the percentage of alumina by removal of water.

Illustrations of copper deposits which have been anamorphosed by rock flowage are rare. The Rammelsberg copper, lead, and zinc sulphides of the Harz Mountains have been deformed by rock flowage, while the associated pyrite has fractured. Pyritic

copper veins at the Milan Mine in New Hampshire [1] have been sheared and faulted and recrystallized while the wall rocks were rendered schistose. There are many copper deposits in schists and associated with sheared intrusives, which may have been formed before rock flowage, as for instance in the Bully Hill district of California or the pyritic copper deposits of Mt. Lyell, Tasmania, but it is difficult to determine in these cases how much of the concentration took place prior to anamorphism. The fact that the ore secondarily impregnates schists and follows cleavage planes indicates that a considerable part has taken place after anamorphism. The Ducktown copper deposits in folded slates of Tennessee are regarded by Emmons [2] as replacements of limestone bands, which have been sheared and recrystallized by rock flowage. Some deformation has clearly taken place since the development of the ore body, but some also doubtless took place during and prior to the deformation. The zinc ores of Ammeberg, Sweden, are disseminated in gneisses resulting from high temperature anamorphism of sedimentary beds.

In the Grand Encampment district of Wyoming there is some evidence of flowage of the copper ores after their formation, but here likewise it is impossible to tell how much of the deposition of the ores took place during or prior to rock flowage.

Concerning the anamorphism of gold and silver ores, there is much the same difficulty as in copper ores. There are many gold and silver deposits associated with schists in which some part of the formation of the ore may have been prior to anamorphism but much of it has been later, as shown by the replacement of the schistose wall rocks. The ore in the Homestake Mine in the sedimentary slates and schists of the Black Hills seems to replace the schists, and yet the evidence is not sufficiently clear to eliminate the possibility that some of it may have been formed before the schistose structure was developed.

It is to be expected that many vein and surface deposits have gone into the zone of rock flowage, but comparatively few have been distinctly recognized as such. This is largely because of the

[1] Emmons, W. H., Some ore deposits in Maine and the Milan Mine, New Hampshire: Bull. 432, U. S. Geol. Survey, 1910, p. 52.

[2] Emmons, W. H., and Laney, F. B., Preliminary report on the mineral deposits of Ducktown, Tennessee: Bull. 470, U. S. Geol. Survey, 1911, pp. 165–168.

fact that during and after rock flowage there have been further changes which tend to mask the evidences of earlier origin.

(3) It is still more difficult to cite specific evidence of the actual segregation of ore *during* dynamic anamorphism. The wide variety of iron ores in regions which have undergone rock flowage shows no evidence that they have been segregating during rock flowage. In fact, the evidence is usually decisive that they were developed under surface conditions and have since been brought into the zone of rock flowage. The Ducktown copper ores, the Encampment copper ores, the Homestake gold deposits, have all been cited as possible cases of segregation of ores during processes of dynamic metamorphism or rock flowage, but the evidence in no case is decisive. There is in all cases the possibility that segregation went on before rock flowage, and certainly since rock flowage a considerable amount of recrystallization and transfer of ore has taken place. Certain zinc deposits in the pre-Cambrian limestones of eastern Ontario are thoroughly recrystallized, lie in bands parallel to the bedding and parallel to the contact of batholithic intrusions, suggesting the possibility that during anamorphism there has been elimination of carbonates, tending to enrich the ore, but more evidence will have to be available to make this any more than a suggestion in the present state of knowledge.

It is ordinarily assumed that under deep-seated conditions of rock flowage by dynamic metamorphism there is so little opportunity for transfers of materials that the composition of the mass is not particularly changed. As compared with the great changes possible in the zone of fracture this is unquestionably true, and yet on the foregoing pages evidences are presented that some changes of composition actually do take place, even on a considerable scale. Water, carbon dioxide, and oxygen are conspicuously eliminated, lime carbonates in some places are replaced by silicates, and even quartzites may lose a considerable amount of silica. These changes suggest the possibility that anamorphic processes may make considerable changes in ores and even segregate ores, but specific inductive evidence from the ores themselves is so indecisive that, as indicated above, it may not be affirmed that anamorphism segregates ores or changes the composition of ores to any considerable extent.

(4) Under thermal and contact metamorphism there has been

considerable segregation of primary ores, both as fillings in fissures and cavities and as replacements of wall rock. The larger number of the vein deposits of gold, silver, and copper, and the replacement deposits of copper, lead, and magnetite along contacts, are distinctly traceable to anamorphic, thermal or contact, conditions. Lindgren [1] has distinguished surface, intermediate and deep groups of deposits, on the basis of the types of minerals and their known temperatures of formations, corroborated in some cases by geologic and physiographic evidence of the depth at which the ore bodies were formed. While these groups are essentially anamorphic in their development, the same solutions which produce them may have distinctly katamorphic effects on the igneous rocks with which they are associated, especially at surface and intermediate depths.

The ore deposits formed near the surface by hot waters in genetic connection with igneous rocks are characterized by gangue of opal, chalcedony, quartz, calcite, aragonite, barite, and chlorite. They contain in places metallic gold and sulphides, such as cinnabar, stibnite, and pyrite, but, according to Lindgren,[2] never the other common ore minerals such as chalcopyrite, galena, zinc blende, and arsenopyrite.

In an intermediate zone, estimated by Lindgren [3] roughly between 4,000 and 12,000 feet, at a temperature somewhere between 175° and 300° C., and with pressures of 140 to 400 atmospheres, there are deposits principally of gold and silver, often with large amounts of copper, lead, and zinc, with predominating gangue of quartz and carbonate. Fluorite and barite are occasionally of importance. Chalcedony and opal are rarely found.

In the deep zone are found cassiterite, wolframite, molybdenite veins, gold-bearing veins and replacements, copper-tourmaline deposits and lead-tourmaline deposits, characterized as a group, with a gangue of pyroxenes, amphiboles, garnets, apatite, ilmenite, magnetite, tourmaline, topaz, micas, spinels, and feldspars. Correlated with these deep zone vein deposits in general intensity of anamorphic conditions are contact metamorphic deposits. The contact deposits have as principal minerals simple oxides and sul-

[1] Lindgren, W., Mineral deposits, New York, 1913.
[2] Op. cit., p. 432.
[3] Op. cit., pp. 513–514.

phides, such as chalcopyrite, pyrite, pyrrhotite, zinc blende, galena, and molybdenite. Magnetite and specularite are the most common oxides. Oxides and sulphides crystallize together. The most common gangue materials are various silicates of lime, magnesium, iron, and aluminum. Among these so-called contact metamorphosed silicates are garnet, epidote, vesuvianite, diopside, tremolite, and wollastonite. Recrystallized and sometimes exceedingly coarse calcite is abundant and quartz is rarely present in large amounts.[1]

Related to the "deep zone" ores formed by hot solutions are primary ores in pegmatite dikes, such as cassiterite, wolframite, columbite, tantalite, monazite, etc.

Where deposits by thermal waters associated with igneous rocks permeate schists, it is often difficult to determine the exact relations of time of ore deposition to rock flowage. In most cases it has clearly followed rock flowage. In some cases rock flowage has undoubtedly been later, but distinctive evidence of it is comparatively rare. The evidences of the succession of the events during the earlier periods of deposition is more or less masked by the later alterations which the ore body has in some cases been subject to.

In summary, cementation enriches ore but anamorphism by rock flowage fails to add much to the value of ores previously formed or to segregate ore bodies though it may recrystallize them and modify their composition. It is essentially a process tending toward complexity of mineral composition, in contrast to the simplifying and segregating effect of katamorphism. Anamorphism by thermal waters in igneous association accomplishes important results in developing ore bodies.

Of great interest to economic geologists is the question whether the ore-depositing waters associated with igneous activity are of magmatic or of meteoric origin. The trend of opinion in recent years has been rather distinctly toward the conclusion that the waters are largely magmatic and that the ores are direct contributions from the magmas, given off during their period of consolidation. Evidence for this is the close association of such ores with igneous rocks both in time and place, the association with high temperature minerals, content of metals, content of flourine, boron,

[1] Op. cit., p. 660.

and other mineralizers, for which a source cannot be easily found other than magmatic, the general mineralogical similarities of ores connected with a given period or province of igneous activity, etc. The evidence falls just short of proof, and it is difficult in any case to eliminate entirely the possibility that meteoric waters rendered hot at igneous contacts have done more or less of the work, particularly near the surface, especially as meteoric waters entirely away from igneous activity are known to accomplish remarkable results in the way of ore concentration. Nevertheless a very high degree of probability has been established for the activity of magmatic waters in producing ore deposits. This subject will not be discussed further, as we are concerned in this book principally with following the successive changes in the metamorphic cycle as expressed in the rocks and ores.

ANAMORPHISM IN RELATION TO NON-METALLIC MINERAL PRODUCTS

The anamorphism of clay tends to make slates and schists. Some of the slates are of commercial value. The anamorphism by rock flowage of a limestone produces coarsely crystalline marbles, which in some cases are of commercial value. Talc slates may be formed by the anamorphism of dolomite. Possibly thermal waters play a part in the process. The anamorphism of a sand may produce a quartzite, which may be of value, but when it is further altered to a quartz-schist it is valueless. Graphite may be formed in association with slates or schists by anamorphism of organic substances. Some occurrences of anthracite are known to be the result of anamorphic processes on bituminous coals and lignites. In general anthracites are found in zones of folding or igneous intrusion. It is clear, however, that other factors enter into their formation, in that anthracite coals are sometimes found in places where there has been neither close folding nor intrusion. Campbell [1] has suggested that the cracking of the cover, allowing escape of gases, is one of the essential factors in the formation of anthracite.

Regional metamorphism of oil shales may have an important

[1] Campbell, M. R., Hypothesis to account for the transformation of vegetable matter into the different grades of coal: Econ. Geology, vol. 1, 1906, pp. 30–33.

surfaces, not necessarily planes. These surfaces are determined by the parallel dimensional arrangement of the mineral constituents, that is, by the mutual parallelism of the greatest, mean, and least dimensional axes of the mineral particles making up the rock mass. They may also be determined by the parallelism of the *mineral cleavages* of the constituent particles.

The Parallel Arrangement of Mineral Particles

A few minerals, such as mica, hornblende, quartz, and feldspar, in various ratios make up all but a very small percentage of schistose or cleavable rocks. To make the discussion concrete, therefore, cleavage will be discussed principally in relation to these four minerals. The technical reader will at once think of qualifications and additions necessary where other minerals are considered, but it is believed that these do not essentially affect conclusions based on the study of a few of the principal schist-forming minerals.

One of the peculiar features of a cleavable rock is the uniformity in shape of the grains of each of the characteristic minerals, determined by their crystal habit. The average ratio of the greatest to the mean dimensions of a mica plate is about 10:1, of hornblende 4:1, and of quartz and feldspar 1.5:1. These ratios are the same whether the rock cleavage is good or poor. In other words, the better rock cleavage does not necessarily mean a greater drawing out or elongation of mineral particles.

When in the laboratory crystals are allowed to develop under stress, they elongate in the plane of easiest relief, supposedly, regardless of habit, but this is not certain, because the experiments have been conducted principally with isometric crystals.[1] Also, crystals not under conditions of growth have been elongated by pressure alone, again more or less regardless of habit. But notwithstanding these experimental results, the minerals in schists have an elongation ordinarily determined by habit alone. The difference between a schist with poor cleavage and one with good cleavage is not so much that the particles of one have been elongated more than the particles of the other, but that it has more of the kinds of particles which by habit are elongated. This mineral form and arrangement in schists is the "crystalloblastic" struc-

[1] Becker, G. F., and Day, A. L., Linear force of growing crystals: Proc. Wash. Acad. Sci., vol. 7, 1905, pp. 283–288.

ture of Becke [1] and Grubenmann,[2] though the writers would qualify their description of this structure. (See pp. 205–207.)

The parallel dimensional arrangement of the mica and hornblende, and sometimes the feldspar, implies a parallelism of their mineral cleavages, because these minerals tend to occur with definite crystal habit within the rock, and the mineral cleavages are definitely oriented with reference to the dimensional axes. The orientation of the dimensional axes of the particles therefore carries with it an orientation of the mineral cleavages. Mica crystals, for instance, lying dimensionally parallel in a schist, have their mineral cleavages in the plane of the two greater dimensional axes, that is, in the plane of rock cleavage. Hornblende crystals lie with their long dimensional axes parallel; the mean or least dimensional axes of hornblende crystals, being so nearly of the same length, may not be parallel. The two cleavages of hornblende are parallel to the major dimensional axes, but are inclined to the minor dimensional axes. Thus the hornblende cleavages in the schistose rocks are parallel to an axis, but not to a plane. The feldspar habit does not give such great dimensional differences. Most of the feldspars in schist show only a slight tendency to assume elongated or tabular shapes due to crystal habit. Their dimensional arrangement is more or less independent of crystallographic arrangement and therefore there is only a slight tendency toward parallelism of the feldspar cleavages.

The dimensional elongation of mica and hornblende parallel to their cleavage faces in schists has been cited as indicating some sort of genetic relationship between elongation and mineral cleavage.[3]

A schistose rock cleaves either between the mineral particles, following the plane of their greatest and mean dimensional axes, or within the mineral particles along their cleavage planes. The first is known as inter-mineral cleavage, and is a capacity to part determined by the dimensional arrangement of mineral particles; the second may be called inter-molecular cleavage, and is related to the ultimate molecular structure of the crystal. Ordinarily when a rock is cleaved the two surfaces show the glistening faces

[1] Becke, F., Über Mineralbestand und Struktur der kristallinischen Schiefer: Compt. Rend. IX Cong. geol. internat., Vienna, 1903, pp. 553 et seq.

[2] Grubenmann, U., Die kristallinen Schiefer, part 1, 1904, part 2, 1907.

[3] Trueman, J. D., The value of certain criteria for the determination of the origin of foliated crystalline rocks: Jour. Geol., Vol. 20, 1912, pp. 236–241.

of hornblende or mica or of other minerals of this type, indicating that the break has followed the mineral cleavages. The parting here has obviously been easier than between the mineral particles. In places where mica and hornblende are not abundant, the cleaved surfaces of the rock show quartz and feldspar, indicating that the breaking has been principally of the inter-mineral type.

Whatever the relative importance of inter-mineral and intermolecular cleavage, it should be remembered that all mineral particles in cleavable rocks are dimensionally arranged, and that this dimensional arrangement involves parallelism of the mineral cleavages only for part of the minerals. Therefore the conclusion is justified that the dimensional parallelism of mineral particles is the controlling factor in rock cleavage; that to this control is due the mutual parallelism of mineral cleavages of mica or hornblende cleavages. Nevertheless it may to some extent be true that the cleavages of these minerals have some influence on their elongation, and therefore on their arrangement. As the dimensions of the minerals of schists are controlled by mineral habit, this becomes an important factor in the structure of schists.

In addition to the dominant features of parallel arrangement above described there is an endless variety of pressure effects expressed in the strained and broken condition of the mineral particles. The straining of a mineral, particularly quartz, modifies the optical properties, with the result that under the microscope between crossed nicols, the crystal shows dark shadows which move across it as the stage is rotated, owing to the directions of extinction differing from point to point. These are known as *strain shadows*. Quartz sometimes shows rows of fluid-filled pores, marking direction of shearing planes, which may be traced through contiguous crystal grains. Micas may yield by shearing movement causing lamellar twinning, parallel to gliding planes, or they may show a bending of crystals. Plagioclase may show a secondary twinning. Orthoclase may develop a *microcline* structure. Pyroxenes, feldspars, and olivine may show *schiller* structures, which are explained by development of cavities, subsequently filled, along easy solution planes, which may be planes of gliding. The breaking of crystals is much more conspicuous in quartz than in feldspar because of its more brittle nature. All stages of cracking and granulation are to be observed, although in

some schists these are absent because recrystallization has healed the fractures. The textures determined by fracture are known in general as *cataclastic* structures. Granulation may be confined to the periphery of grains or groups of grains, leaving the cores unbroken. This is sometimes called the *mortar* structure, or the *augen* structure. The complete granulation of constituents is sometimes expressed by the term *mylonite*, meaning a rock which has been milled out, although this term in actual usage covers also schists in which evidence of recrystallization is important. The crystals may be broken along parallel planes and the parts relatively displaced, as the result of *slicing*.

Manner in Which the Parallel Arrangement of Minerals is Brought About

The arrangement of the mineral constituents of a cleavable rock is the result of the differential pressure, which caused the rock to flow. Briefly, the general conditions of rock flowage have been found to be great pressure from all sides, high temperature, abundance of altering solutions, susceptibility of the rock to mineral and chemical change depending on its composition, and slow deformation; in other words, rock flowage, judging from the field and laboratory evidence, is accomplished by means of physical and chemical changes combined. These general observations do not indicate just in what manner the parallel arrangement of mineral constituents producing the cleavage, the most conspicuous result of rock flowage, has been accomplished.

The constituents of the rock being diverse, different parts move with different ease and in diverse manners, so that differential flow ensues, some parts moving faster than others. Some parts may recrystallize while others granulate. Differences in ease and amount of flow to be observed between the different mineral constituents of the schist are similar in all respects to the differences shown in a larger way by interstratified rock bodies of far differing strength and character, like limestone, sandstone, shale, or igneous masses.

Recrystallization. A study of cleavable rocks shows that much of the hornblende and mica, minerals which are responsible for some of the best rock cleavage, is of entirely new generation in the secondary rock. A clay may have no mica; a phyllite, its altered

equivalent, may have as high as 50%, by weight, of mica. Chemical analysis shows that this change may occur in some instances with little addition or subtraction of materials. The inference is that the new minerals of the hornblende and mica types have developed principally from the recrystallization of substances already in the rock mass. Even where there is quantitative evidence that substances have been introduced or extracted, the mass has still been recrystallized. Since hornblende and mica are the common minerals producing the best rock cleavage, it must be concluded that recrystallization is the important process in the development of parallelism of the mineral constituents.

Corroborative evidence of the importance of recrystallization is the general lack of fractures or other strain effects in the minerals of a cleavable rock, such as would be expected if the parallelism had been brought about entirely or largely by mechanical processes. It may be inferred, then, that some constructive process, which may be called generally recrystallization, has been at work.

Most of the mineral particles in the cleavable rocks are individually larger than the particles in the same rocks before flowage had occurred. For instance, the gradation of a shale to a phyllite means an increase in the size of the grains. Recrystallization is the constructive process which has accomplished this result.

The cleavable rock is likely to show a great uniformity in size and shape of the grains of the same mineral as compared with the non-schistose rock, and again recrystallization explains the phenomenon.

Much detailed microscopical evidence might be cited, such as dove-tailing of quartz individuals in quartz bands, the feathering out of mica plates against an adjacent mineral surface, the lack of bending and breaking of hornblende needles by mutual interference, the segregation of minerals into bands, to show that the parallelism could not have been produced by mechanical adjustment alone, but must have been aided by the chemical and mineralogical changes involved in recrystallization.

Granulation and rotation of original particles. Recrystallization is not the only process instrumental in the production of rock cleavage. The quartz and feldspar in the cleavable rock may be largely original quartz and feldspar; some of the mica and hornblende also may be original. Parallelism may be partly due to

rotation from original random positions. This process may be aided by granulation and slicing of the original mineral particles. Broken, unequidimensional mineral fragments are often strewn out in such a manner that their longer dimensions lie approximately parallel. Evidence of rotation is seen principally in the quartz and feldspar, which have not much effect in producing rock cleavage. It is concluded, then, that the rotation of original particles, diversely oriented, to a parallel position is a minor factor quite subordinate to the dominant process of recrystallization.

The abundant muscovite developed in the paste or matrix of shales, prior to rock flowage, does not show parallel arrangement, but in the slates resulting from the dynamic metamorphism of these shales, muscovite has a parallel arrangement. Presumably rotation of the muscovite in the shales is a factor in this process. Yet, muscovite in the slates is more abundant, in larger flakes, and more cleanly separated from the chlorite, suggesting that here also recrystallization has played its part.

In the incipient stages of rock flowage the larger and more brittle particles are granulated and elongated. At the same time recrystallization, beginning on the finer particles, builds up new a few minerals like calcite, which are not important in cleavable dominates over granulation and ultimately obliterates any evidence of it. It may be inferred that granulation aids recrystallization in that it grinds the particles into small pieces and affords greater surface upon which the chemical process may act.

In experimental deformation the conditions are not favorable for recrystallization, and granulation is the important process.

Slipping or twinning along the cleavage planes of minerals, called "gliding"—such as may be observed in calcite and ice crystals—has been cited as a possible cause of the elongation and parallel arrangement of mineral particles. This has been observed only in a few minerals like calcite, which are not important in cleavable rocks; and even in the calcite of schistose rocks gliding has been found to be subordinate to processes of recrystallization and granulation. In experimental deformation of marble it seems to play a greater part, because conditions of recrystallization are not present.

There is no evidence that the flattening of original mineral particles to a dimensional parallelism, without regard to crystallographic arrangement, has played any important part in the pro-

duction of rock cleavage; indeed, some of the facts already cited constitute decisive evidence to the contrary. Such is the evidence that hornblende and mica, essential minerals of schistose rocks, are in many cases, and perhaps in most cases, entirely new developments in the rock. Of the same nature is the evidence derived from the uniformity of dimensional characteristics of the particles of a given mineral species and the control of dimensions by crystal habit. The most cleavable rock is not made up of flatter particles of hornblende, mica, quartz, or feldspar than the less cleavable rock. But it certainly contains more particles of hornblende and mica than of quartz and feldspar; consequently it has more particles which are flat or elongate, which give it a better and smoother cleavage.

If this is true, the development of rock cleavage would seem to require change in chemical composition necessary to increase the proportion of the cleavage-making minerals, such as hornblende or mica. Chemical evidence seems to the writers to point this way, though it is not yet sufficient for proof.

Cleavage in its Relations to Differential Pressures

It has been shown that rock cleavage is determined by the parallelism of mineral constituents and that this parallelism is developed by rock flowage, which implies differential pressures sufficient to cause movement. It now remains to discuss the attitude of cleavage with reference to specific pressure conditions.

What experimental evidence there is indicates that in a non-rotational strain [1] mineral particles tend to arrange themselves with their longest dimensions normal to the direction of the pressure. There is practically no experimental evidence bearing on the arrangement of particles under rotational strain or shearing, so common in nature.

Wright [2] melted about 50 grams each of wollastonite, diopside, and anorthite, and plunged the melt into water, thereby forming a glass. Cubes were then cut from these glasses, heated to a viscous state at which crystallization first begins, and subjected to vertical pressure. Microscopic examination showed that the three minerals

[1] Leith, C. K., Structural Geology, New York, 1913, pp. 16–21.

[2] Wright, F. E., Schistosity by crystallization. A qualitative proof: Am. Jour. Sci., vol. 22, 4th ser., 1906, p. 226.

named had crystallized with their longer dimensional axes normal to the pressure.

Becker and Day [1] have shown that although crystals are able to grow in a given direction in spite of contracting forces, their growth in the plane normal to the pressure is vastly greater, whether this be the normal direction of elongation due to habit or not. Ordinarily in schists the elongation of the crystal is that of its normal habit, indicating perhaps that the crystals favorably oriented to grow with normal habit have grown at the expense of those not favorably oriented.

Field observations have to do principally with the relation of cleavage to the elongation of the rock mass (strain) which can be seen, and not with *stress*, which cannot be seen and may only be inferred from the strain. After having proved the relation of cleavage to strain, the general relations of strain to stress may be considered.

It seems self-evident that the longer dimensions of mineral particles in a cleavable rock lie parallel to the elongation of the rock mass developed during rock flowage. This relationship has been so generally assumed by geologists that at first thought it would seem entirely superfluous to present evidence in proof of it. But it has been questioned by able geologists. Becker [2] has held that the elongation of the rock mass may be inclined to the common direction of the major axes of the mineral particles. The student, when asked how he knows that cleavage is parallel to rock elongation, is often completely at sea. It is simply a matter of observation to determine definitely whether the cleavage is parallel to the elongation of the mass as a whole. Evidence indicating this parallelism is as follows: (1) *Distortion of pebbles of a conglomerate.* Schistose conglomerates show by the distortion of their pebbles the plane of elongation, although it may sometimes be difficult to distinguish the shapes of undeformed pebbles from those of deformed ones. The cleavage of the matrix is approximately parallel to the greater diameters of the flattened pebbles, although it curves somewhat at the ends of the pebbles. (2) *Distortion of*

[1] Becker, G. F., and Day, A. L., The linear force of growing crystals: Proc. Wash. Acad. Sci., vol. 7, 1905, pp. 283–288.

[2] Becker, G. F., Current theories of slaty cleavage: Am. Jour. Sci., vol. 24, 4th ser., 1907, pp. 1–10.

mineral crystals. The plane of cleavage is marked by mica plates or hornblende crystals, while the associated quartz and feldspar particles may be fractured at angles with the plane of cleavage. The displacement of the parts, which often accompanies such fractures, is observed to extend the fractured parts in the plane of rock cleavage. (3) *Distortion of volcanic textures.* The original ellipsoidal parting of basalts frequently shows a flattening, with or without fracture; in such cases the ellipsoids and the matrix have a flow cleavage parallel to the longer diameters. The elongation of amygdules and spherulites in planes parallel to the rock cleavage is likewise of common occurrence. (4) *Distortion of fossils.* The elongation of fossils in the plane of cleavage has been observed in cleavable rocks. (5) *Distortion of beds and attitude of folds.* Folds often show the direction of shortening of the deformed rock mass. (6) *Relations to intrusives.* Intrusions of great masses of igneous rocks, and particularly deep-seated batholiths, exert pressure against their walls. Any cleavage developed in the surrounding rocks is parallel to the periphery of the intrusive masses.

It is concluded then that the longer dimensions of mineral constituents are parallel to the directions or planes of elongation of the rock mass. Thus an adequate statement of the relations or rock cleavage to the stresses which have produced it must be a statement which will cover the various ways in which stress has elongated and shortened rock masses.

In the simplest possible terms stress has been effective in distorting rock masses: (1) by non-rotational strain (pure shortening) in which the axes of stress and strain remain mutually constant throughout the deformation, and (2) by rotational strain (shearing) in which there is a continuous change in the position of the strain axes as compared with the stress axes during the distortion.[1] In the first case the elongation of the rock mass is normal to the greatest stress and remains so through the deformation; in the second case the elongation of the rock mass is constantly changing in direction with reference to the principal stress, and ultimately the elongation may be considerably inclined to the maximum stress.

Substituting rock cleavage for greatest elongation of the rock mass, the statement of the relations of cleavage to pressure is as

[1] Leith, C. K., Structural geology, pp. 16–21, New York, 1913.

follows: In a non-rotational strain cleavage is developed normal to the greatest stress; in rotational strain, while at any instant there may be a tendency for it to be developed normal to the greatest stress, there is here a rotational element which brings it into position inclined to the greatest stress. All distortional strains in rock masses belong to these two classes, rotational and non-rotational, and usually to some combination of the two. Cleavage, therefore, is developed under some combination of rotational and non-rotational strains and may be said to be produced both normal and inclined to pressures.

This relation is often misstated in text books, and it is a source of much confusion in field work. It is not adequate to say that cleavage is developed normal to the pressure. It is developed *parallel to the elongation of the rock mass*, which in turn may or may not have been developed normal to the greatest pressure. Careful observation will frequently tell which of the two cases may be under observation, and if so, the cleavage may become of much significance in inferring other structural conditions.

There is field evidence to show that cleavage or schistosity develop only when there has been actual movement in the rock. The use of the term "dynamic" in describing this type of metamorphism implies belief in movement as essential to its production. Where the differential stresses are not sufficient to cause movement, there is no satisfactory evidence that the schistose or cleavage structure has been produced. In other words, there must be not only differential stresses but the difference in stresses must be great enough actually to cause movement, before schistosity or cleavage develops. Great pressures at depth are not necessarily adequate to cause rock flowage. There must be sufficient stress *differences* to overcome resistance. With greater depth rigidity or resistance increases, and it is even possible that this generally surpasses the stress differences, in other words, that there is no zone of flow at a given depth where all rocks must flow.

GNEISSIC STRUCTURE

Gneissic structure means a banding of constituents, of which feldspar is important, with or without the parallel dimensional arrangement necessary for rock cleavage. A schist always has a

parallel dimensional arrangement and may or may not contain feldspar. A gneiss may or may not have a parallel arrangement, but always has a banding and contains feldspar. So far as this parallel arrangement is present, gneissic structure has been discussed under the heading of rock cleavage. In many cases, however, cleavage in gneisses is not good. The essential mineralogical difference between gneisses and schists is the possession by the gneisses of a relatively small amount of the platy and columnar minerals so necessary for a good rock cleavage, and correspondingly more feldspar and quartz. Differences in origin are discussed on pages 153 et seq.

IDIOMORPHIC OR PORPHYRITIC TEXTURES DEVELOPED BY ROCK FLOWAGE

Garnet, staurolite, tourmaline, andalusite, chloritoid, and other heavy anhydrous minerals of this kind are uniformly idiomorphic or porphyritic in cleavable rocks. They develop by recrystallization after rock flowage has ceased, but probably while the rock is still under high pressure and temperature, as is evidenced by their high specific gravity and frequent occurrence in the proximity of intrusive igneous rocks. Their late development by recrystallization is shown by the following considerations: (1) They appear in rocks clearly derived by rock flowage from others originally lacking such minerals. (2) They frequently lie at large angles to the prevailing cleavage in the rock. (3) They do not show the degree of mechanical deformation that they would necessarily have possessed had they developed in their present positions before flowage had ceased. Many of the crystals are long and acicular, and would surely have been broken if any considerable movement had occurred subsequent to their development. (4) They include, within their boundaries, minerals in part similar to those in the remainder of the rock, and which have an arrangement of their greater diameters in the plane of rock cleavage, showing that to some degree at least such minerals were formed during rock flowage and that the porphyritic developments came later. (5) The mica and the other constituents of cleavable rocks, which are certainly developed by recrystallization during the process of deformation, are frequently seen to end abruptly at the periphery of a mineral of this group and not

to curve around it as they often do about the resistant minerals in schists. If the rock had flowed after the formation of the porphyritic crystals, crowding and bending of the micas must inevitably have occurred. (6) The usual large size of minerals of this group, as compared with their associated mineral particles, suggests their development subsequent to rock flowage, when granulation is no longer tending to break down the crystals.

While the development of this group of crystals is believed to have been mainly later than the formation of the cleavage, it is true also that in some cases subsequent flowage has resulted in their being fractured and crowding the other constituents. The very fact that the effects of further movement are so conspicuous confirms the conclusion that the secondary porphyritic minerals not showing these effects developed after the movement ceased.

We may only speculate as to the conditions of this peculiar development of non-arranged minerals. They are probably those of high temperature and pressure, but apparently not of movement. If the pressure and temperature may be considered as having become so great as to develop hydrostatic conditions, there would be no differential pressures necessary for a parallel arrangement of constituents and this might afford a plausible explanation of the development of these non-oriented porphyritic minerals.

Another possible explanation is suggested by the experimental fact that rocks under stress, but with lateral support, take on a great rigidity, which makes it possible for them to resist considerable stress differences without yielding. It seems entirely possible that the schists, at the time of development of the porphyritic constituents, may have been under great differential stresses, but because of the high rigidity, due to lateral support, may have been able to withstand these stresses and not yield further by rock flowage. It may be supposed that rock flowage, developing schistosity, tended toward equilibrium of differential stress conditions, thus lowering the stress difference to a point where it was not great enough, especially with considerable rigidity in the rock mass, to cause movement. Under such conditions, it would seem possible for porphyritic constituents to develop. These conditions would differ from hydrostatic conditions in that there would still be considerable differential stress, but not sufficient to cause movement.

Whatever the explanation, the uniformly high density of the porphyritic constituents suggests that they have formed in response to demand for less volume.

ROCK FLOWAGE WITHOUT RETENTION OF CLEAVAGE

Marble is the commonest example of a rock which undergoes flowage without retaining cleavage. It often occurs between schistose beds which have flowed, without doubt the marble itself has flowed, and yet it possesses no cleavage. Cleavage may be produced experimentally in marble by pressure alone, when the conditions are not favorable for recrystallization.[1] Microscopic examination indicates that this has been accomplished by granulation, slicing, and gliding of the calcite crystals. Rarely such a cleavage is observed in marble deformed under natural conditions. It may be supposed that many marbles have shown this structure in the early stages of their flowage, but calcite recrystallizes so easily that the parallel structure caused by mechanical deformation is soon destroyed. The recrystallized calcite crystals do not have the habit necessary for a good dimensional arrangement in schists.

So far as the limestones have impurities in them, secondary silicates are likely to develop, such as actinolite and tremolite, which by their arrangement may give the rock a cleavage.

OBLITERATION OF PRIMARY TEXTURES BY ROCK FLOWAGE

Recrystallization, the dominant process in rock flowage, tends toward an increase in the size of grain, the segregation of minerals into bands, a uniformity in size and shape of the mineral particles, and the growth of new minerals such as mica or hornblende not previously existent in the rock. Previous textures are commonly destroyed. Bedding is locally not completely obliterated, because alternation of beds of originally different mineralogic character and texture determines to some extent the kinds and size of the secondary mineral particles formed in these beds by rock flowage. Thus a faint banding of dark or light minerals or of fine or coarse minerals may mark the original bedding in a schistose rock.

[1] Adams, F. D., and Nicolson, J. T., An experimental investigation into the flow of marble: Phil. Trans. Roy. Soc. of London, vol. 195, 1901, pp. 363–401. See also Adams, F. D., and Coker, E. G., The flow of marble: Am. Jour. Sci., vol. 29, 1910, pp. 465–487.

CHAPTER VII

CONDITIONS OF DEVELOPMENT OF CRYSTALLINE SCHISTS AS CONCEIVED BY BECKE AND GRUBENMANN

While no attempt is made in this volume to discuss comprehensively the voluminous literature on the origin of the crystalline schists, reference may be made to the well known views of Becke and Grubenmann.

Becke [1] and Grubenmann [2] have discussed the crystalline schists in relation particularly to the molecular volumes and heats of formation of the constituent minerals, and draw certain inferences as to factors and conditions controlling their development.

BECKE ON THE CRYSTALLINE SCHISTS

Becke points out the difference in mineral composition between the igneous rocks and the crystalline schists and infers that this difference is due to the fact that the minerals in the igneous rocks crystallize at high temperature, while those in the schists form at lower temperature. The development of igneous rocks involves differentiation; the development of crystalline schists tends toward uniform condition of chemical equilibrium. Differences in composition are wiped out. Homegeneous isomorphous mixtures of minerals are developed and exist at the higher temperatures, but at lower temperatures these break up into heterogeneous more or less mechanical mixtures.

Following Loewinson-Lessing's [3] method of calculation, the molecular volume of the common minerals of igneous rocks and

[1] Becke, F., Über Mineralbestand und Struktur der kristallinischen Schiefer, Compt. Rend. IX Cong. geol. internat., Vienna, 1903, pp. 553 et seq.

[2] Grubenmann, U., Die kristallinen Schiefer, vol. 1, 1904, vol. 2, 1907.

[3] Loewinson-Lessing, F., Studien über die Eruptivgesteine: Compt. rend. VII Cong. geol. internat., St. Petersburg, 1897, pp. 194 et seq.

schists are compared with the sums of the molecular volumes of their constituent oxides. When the molecular volumes of the minerals are greater than that of the constituent oxides they are called *plus* minerals, when less they are called *minus* minerals. In comparing the distribution of the plus and minus minerals, it is found that most of the minerals which play an important rôle in igneous rocks and contact rocks, but which are lacking in the normal crystalline schists, are plus minerals, while those in the crystalline schists are stated to be minus minerals. There are, however, important exceptions, as in the case of the feldspars, which are plus minerals and are found in the crystalline schists. Inspection of the table of plus and minus minerals shows so many exceptions as to raise the question as to whether this distinction is a valid one.

It is argued that the minus minerals are developed and controlled by pressure and that the plus minerals are controlled in their development by heat, which acts against and withstands the pressure.

A number of the common minerals of igneous rocks are compared with supposedly equivalent minerals in the crystalline schists in order to show that there has been a diminution in molecular volume. These are based on comparison of minerals of similar or identical composition, but they are not based strictly on chemical reactions of definitely observed alteration; for instance, augite is compared with garnet, augite and anorthite together are compared with garnet, and quartz, olivine, and anorthite, with garnet. No attempt is made to calculate the molecular volume of the rock as a whole, or even to select minerals of identical composition.

The writers doubt very much the validity of Becke's generalization that decrease in molecular volume in the development of schists has been established. In fact, Becke states that the deductions from the volume relation should not be carried too far.

It is interesting to note how Loewinson-Lessing's tentative suggestion, that molecular volumes are a factor in determining the order of crystallization of plutonic igneous rocks, has been extended by Becke to the consideration of schists. Loewinson-Lessing, considering only the plutonic igneous rocks, concluded that minerals with a less molecular volume than that of their constituent oxides are on the whole the first to crystallize, but this

was scarcely more than a suggestion, for he stated that the future would show whether this apparent rule really holds and whether it can be brought into relationship with thermal chemistry and the solution theory. It would seem that the idea had not found much favor, for it is not even mentioned in well known discussions of magmatic differentiation, such as those which appear in Harker's, Idding's, and Daly's textbooks of petrology.

Becke extends the idea to the schists and concludes that the igneous rocks, on the whole, are characterized by plus minerals and the schists by minus minerals. Grubenmann, whose work is summarized on a later page, carries the idea still further, and, in a more detailed classification of the schists, finds two upper zones characterized by minus minerals and a lower one by plus minerals. All three investigators conclude that so far as this "law" of molecular volume does not hold, it is because of heat changes in the reactions, indicating temperature control. Milch [1] essentially follows Grubenmann, but divides the schists in two zones instead of three.

If it were proved that the molecular volumes of the minerals of the schist were, on the whole, less than those of igneous rocks, there is still doubt as to the significance of this in relation to pressure and temperature control. Certain it is that both plus and minus minerals may develop side by side under a great variety of conditions. It is difficult to say that one group of minerals is determined by pressure and another by temperature. This is true not only in the crystalline schists but in the products of weathering. For instance, kaolin is a minus mineral, halloysite is a plus mineral, gibbsite is a minus mineral, and bauxite is a plus mineral. In fact, so far as the writers know, no satisfactory law has been developed which is demonstrated to control the molecular volumes of the minerals developed. The relation of molecular volumes of minerals to the sum of the molecular volumes of their constituent oxides seems to be only one of the various properties of minerals, and certainly the other properties, particularly composition, habit, and density, play an important part in determining the development of a mineral under any given environment.

Physical chemists are unable to tell why the molecular volume

[1] Milch, L., Die heutigen Ansichten über Wesen und Enstehung der kristallinen Schiefer: Geol. Rundschau, vol. 1, 1910.

of a mineral is greater or less than that of its constituent oxides. Even in putting two solutions together they are unable to predict whether the molecular volume will be increased or decreased. In fact, some of them are skeptical about the actual determinations of molecular volume, due to the considerable variation in density of solids of a given composition and lack of information as to the real molecular composition of solids. It is difficult, for instance, to determine what specific gravity to use for the silica, as this oxide takes on a number of crystalline forms with differing densities. Also the molecular weights of solids are problematical, as no means of exact determination have been found. The molecular weights are determined principally from gases and liquid solutions. The molecular weight of silica in the solid form is not definitely known. The relative weights of silicon and oxygen can be determined, but whether the molecule is made up of one or more SiO_2 units in the solid is not known. For that matter, the molecular weights even in liquid and gaseous solutions are variable; for instance, the molecular constitution of sulphur in gaseous form ranges from S to S_6, with changing temperatures and in solutions of carbon bisulphide or napthalene it is S_8. Finally, the very existence of oxides in magmas and solutions has been questioned.

Crystalloblastic Structure

Becke distinguishes between characteristics of textures of igneous rocks and those of schists. The texture of igneous rock is controlled by successive crystallization of minerals. In crystalline schist there is no successive crystallization. There is developed a so-called *crystalloblastic* texture, the characteristics of which are as follows: (1) The constituents are of equal rank and show no definite order of crystallization. Any one mineral may occasionally exist as an inclusion within another. The later formed minerals may have more perfect crystal outlines than their inclusions. (2) Crystal forms are rare. Those present are usually simple, and there is marked development parallel to the cleavage planes. (3) Lack of skeleton crystals. (4) The minerals in the schists with the most compact molecular arrangement are most likely to take on crystal form, that is, have the strongest crystallizing power, and on this basis Becke develops a series of "crystalloblastic" minerals named in order of their decreasing crystallizing power.

The usual series of form development is as follows: titanite, rutile, hematite, ilmenite, garnet, tourmaline, staurolite, cyanite-epidote, zoisite-pyroxene, hornblende-magnesite, dolomite, albite, mica, chlorite, talc-calcite-quartz, plagioclase-orthoclase, microcline. (5) Parallel structure of the crystalline schists is the result of rotation of already existing minerals and of recrystallization. (6) Crystalloblastic structure is holocrystalline, not cellular. (7) Zonal structure in minerals is absent, or, if present, follows different rules than in igneous rocks. (8) Inclusions do not follow the strata of the crystal but are related to the pyramids added to original or else they are related to an older helizitic structure.

From our standpoint this statement of the principal features of secondary gneissic and schistose textures is essentially correct, but we would suggest certain qualifications, principally in regard to the statement that the minerals in schists develop in a crystalloblastic order determined by their crystallizing power and density. In general, we agree to the conclusion that minerals in schists, with the best crystal development have the strongest crystallizing power and are in general dense, but that this is due to their crystallizing power or to their density alone does not necessarily follow. As indicated on pages 204–205, during the movement of the rock mass the development of new minerals seems to be influenced by crystal habit or dimensions, for during this phase of the development of schists and gneisses only columnar and platy minerals are developed, adapted dimensionally to the conditions of movement required by the unequal pressures. Later there may be porphyritic development of garnet, tourmaline, staurolite, and other heavy anhydrous minerals, arranged entirely independent of the schistosity, indicating that there has been no movement during or subsequent to their development, and that they have probably developed under mass static conditions. The high average density of these minerals suggests that density also may have been a factor in their development.

We would question also the statement that the mineral constituents of the schists are of equal rank and show no definite order of crystallization. The order seems to us a fairly definite one: first, the development of platy and columnar minerals during rock movement, later, the development of porphyritic anhydrous constituents, where there is no movement. Each mineral newly

developing in a schist derives its material from the destruction of preëxisting minerals. This differs from the conditions in the igneous rocks, where minerals crystallize from a magma, and those with the strongest crystallizing power presumably crystallize first, the remainder filling the spaces between.

Finally the statement may be questioned that crystal forms are rare. The characteristic development of platy, columnar and porphyritic minerals in the schists, with their own distinctive habits, means a uniform approach to crystal form, which does not seem to the writers to be less marked than in igneous rocks, where many of the later developments fill the spaces between the early developed crystals, and, therefore, cannot take on their own crystal form. So far as the writers' observations and measurements go, the tendency is for the common minerals of the schist to take on crystal habits not far different from those in the igneous rocks.[1] Trueman[2] noted a similarity in shapes of crystals in schists and igneous rocks, but concluded from measurements that on the whole the crystals in the schists were somewhat more elongate than those in the igneous rocks.

GRUBENMANN ON THE CRYSTALLINE SCHISTS[3]

Grubenmann proposes a detailed classification of the schists and gneisses in relation to their locus and conditions of development. He argues that schists and gneisses are produced in three zones, each with its distinctive physical conditions and resulting mineral products, and follows Loewinson-Lessing and Becke in emphasizing consideration of molecular volumes as an important factor in determining development of minerals in crystalline schists.

The conditions named for the upper zone are moderate temperature and low hydrostatic pressure, positive heat-toning,[4] and

[1] See Leith, C. K., Rock cleavage: Bull. 239, U. S. Geol. Survey, 1905, pp. 24–48.

[2] Trueman, J. D., The value of certain criteria for the determination of the origin of foliated crystalline rocks: Jour. Geol., vol. 20, 1912, pp. 236–241.

[3] Grubenmann, U., Die kristallinen Schiefer, vol. 1, 1904, vol. 2, 1907, Zurich.

[4] The "heat-toning" of a reaction is positive when the sum of the resulting heats of formation exceeds the sum of the heats of formation of the vanished molecule. Positive heat-toning is exothermic and negative heat-toning is endothermic.

powerful stress. The predominant pressure effect is mechanical. Heat liberation and decrease of volume result from the chemical reactions. The abundance of water favors the development of hydroxide minerals, such as sericite, chlorite, talc, zoisite, and epidote. These minerals, with quartz and calcite, form the characteristic rocks of this zone, namely, sericite-phyllite, sericite-quartzite, lime-phyllite, albite-phyllite, talc, chlorite-schists, etc.

The conditions named for the middle zone are higher temperature and intense stress, stronger hydrostatic pressure, and positive and negative heat-toning. The predominant effect of pressure is recrystallization and remineralization with decrease in volume. The minerals assigned to this zone as characteristic are biotite, zoisite, epidote, hornblende, staurolite, garnet, disthene, titanite, magnetite, ilmenite, muscovite, microcline, albite, and oligoclase. The typical rocks of this zone constitute the great mass of the crystalline schists: mica-schists, amphibolites, garnet, and staurolite-schists.

The specified conditions of the lowest zone are enormous hydrostatic pressure and high temperature, very feeble stress, and negative heat-toning. The reactions are inferred to take place with volume increase, since the minerals of this zone nearly all belong to Loewinson-Lessing's "plus" group. ("The molecular volume of the minerals of the plus group exceeds the molecular volumes of the oxides which compose them.") Grubenmann argues from this that the reactions are probably controlled by temperature and not by pressure. The minerals which are described as peculiar to this zone are cordierite, magnetite, sillimanite, ilmenite, orthoclase, plagioclase, biotite, augite, olivine, and garnet. The gneisses constitute the predominant rocks.

A number of mineral and rock alterations are cited as peculiar to each zone. Olivine of the lowest zone becomes hornblende or garnet in the middle, and serpentine in the upper zone. Granulations of grain, chloritization, and sericitization occur in the upper zone. An aluminous sediment composed of kaolinite, quartz, mica, feldspar, iron oxide, and lime becomes a phyllite in the upper zone, a mica schist in the middle, and a gneiss containing feldspar, sillimanite, and other anhydrous minerals, in the lower zone.

Grubenmann then divides the schists into twelve groups de-

limited by empirically derived chemical values. The groups take their nomenclature from the rock formations which are peculiar to them. They are (I) orthoclase gneiss, (II) alumina-silicate gneisses, (III) plagioclase gneisses, (IV) eklogite and amphibolite, (V) magnesian-silicate gneisses, (VI) jadeite rocks, (VII) chloromeianite rocks, (VIII) quartzitic rocks, (IX) lime-silicate rocks, (X) marbles, (XI) iron-oxide rocks, and (XII) alumina-oxide rocks. Each group is subdivided into orders, which are based on the physical characteristics assumed by a group in the upper, middle, and lower zones respectively. A group-name takes the prefix "kata" when it designates a rock of the first order or of the lowest zone; the prefix "meso" is applied to rocks of the middle zone or the second order; and the rocks of the third order or the upper zone take the prefix "epi." Frequently the orders are subdivided into families, based on mineral compositions.

While Grubenmann has brought together much valuable information on the schists, the writers are unable to follow him fully in his general classification and the inferences he draws from it. Their principal objection is to his classification of the schists and gneisses into upper, middle, and lower zones. It puts an emphasis on depth as a controlling factor in anamorphism which does not seem to the writers to correspond to actual field observations. So far as the upper, middle, and lower zones represent successively more intense conditions of anamorphism there is some basis for the classification, but depth is only one of the factors determining these differences in intensity. Local dynamic metamorphism or igneous intrusion may accomplish the most intense changes at moderate depth, while in the absence of certain favorable conditions of this kind certain rocks may be deeply buried and yet undergo but slight anamorphism. A competent formation like a quartzite may require very great depth and extreme conditions of anamorphism to change it along local zones to a sericite schist, and yet a sericite schist is classified by Grubenmann as belonging to an upper zone of slight metamorphism. As important as depth are physical competency of rocks to withstand rock flowage, their mineral and chemical composition determining the ease of their recrystallization and rock flowage, their water content, the presence or absence of localized mechanical shears, presence or absence of intrusives, temperatures, conditions allowing free escape of sub-

stances eliminated by anamorphism, such as water, carbon dioxide, and oxygen, and probably other factors. Because of this variety of determining conditions, the results of rock flowage at a given depth may be highly variable. While it is possible to group schists and gneisses on the basis of general intensity of the conditions they represent, even this is not always safe, for field conditions indicate that the same conditions which will produce certain schists of the "upper" group from one type of rock, may produce schists of the "middle" or "lower" groups from another type of rock. For instance, a chlorite schist may form by the anamorphism of basic sediment or basalt under the same conditions that amphibolites may be the result of the anamorphism of a limestone.

Grubenmann does not distinguish between dynamic and contact metamorphism in the production of groups of the several zones. By including augite and olivine in his lower zone, one may suspect that he is including in his schists and gneisses slightly schistose igneous rocks in which the minerals are of primary igneous origin, or igneous rocks like the primary or protogene gneisses (see p. 153), for certainly neither olivine nor augite is a characteristic secondary product of the dynamic and contact processes known to produce schists and gneisses. In fact, Grubenmann says, in describing augite schists and olivine schists, that these in most cases show but a slight amount of schistosity and that there is difficulty of discriminating them from primary igneous rocks. While there doubtless are metamorphic schists and gneisses in which augite is developed secondarily, the writers' view is that augite is so rare that it can scarcely be used as a basis for classification of schists. For identifying the several zones he uses minerals which are distinctly characteristic of dynamic metamorphism, some characteristic of contact metamorphism and some even characteristic of igneous rocks. While these may represent general stages of intensity, they do not necessarily represent depth.

Grubenmann infers lessening of the volume in the two upper zones and increase in the lower zone. In the upper and middle zones the molecular volume of the principal minerals is less than the molecular volume of the constituent oxides, therefore, decrease in volume is inferred. In the lower zone the molecular volumes of the principal minerals are greater than the sum of molecular volumes of their constituent oxides; therefore, increase

in volume is inferred. In inferring an increase in volume for the lower zone, Grubenmann departs somewhat from the generalization of Becke that the schists as a whole show a decrease of molecular volume as compared with the igneous rocks. This suggests that Grubenmann regards the conditions of his lower zone as being essentially those favorable to the development of igneous rocks.

Grubenmann's method of calculation of volume does not take into account changes in volume by the elimination of pore space, or the elimination of water, carbon dioxide, and oxygen, or other substances. It does not take into account the volume of the minerals actually destroyed. As already noted on a preceding page, it is very doubtful whether emphasis on molecular volumes is warranted. This property of the minerals is only one of several properties, including crystal habit, composition, and density, which are known to have some influence in determining the development of crystals in a given location.

Until far more laboratory data are available conclusions as to the net results of the heat reactions and energy changes must be tentative. Grubenmann does not give data on which he bases his conclusion that the chemical changes in his upper zone are exothermic and in the lower zone endothermic, nor in looking over the list of minerals which he cites as characteristic of these zones do we find that the heats of formation have in many cases been positively determined. The hydrous character of some of the minerals of the upper zone may have suggested an exothermic reaction, but so far as these minerals are the result of dehydration of more hydrous minerals the heat change may have been the reverse. For instance, sericite might develop from feldspar with an exothermic heat change, but so far as it developed from colloidal clay and potash it is probably endothermic. Grubenmann regards his view of heat changes as somewhat similar to that of Van Hise,[1] who argued liberation of heat by chemical reactions in an upper zone of katamorphism and absorption of heat in general by the chemical reactions in a lower zone of anamorphism. These two zones, however, should not be compared with the three schist zones of Grubenmann, as Van Hise was considering both katamorphism and anamorphism, and Grubenmann was consider-

[1] Van Hise, C. R., A treatise on metamorphism: Mon. 47, U. S. Geol. Survey, 1904, p. 182.

ing different phases of anamorphism, producing schists. It is our view (essentially following Van Hise) that anamorphic changes in general mean absorption of heat so far as the chemical reactions are concerned; some of them are known to be endothermic and the probability is that most of them are endothermic. Considering the similarity of conditions and results of anamorphism it would seem to us doubtful that it includes sharply differentiated groups of reactions of both endothermic and exothermic kinds. The most definite evidences of heat changes are to be found in the simpler katamorphic changes. These are mainly exothermic. There is abundant proof that anamorphic changes are in most essential respects the reverse of those of katamorphism, which suggests that the heat changes are also the reverse, in other words, endothermic.

CHAPTER VIII

GENERAL CONSIDERATION OF ANAMORPHISM

TEXTURES

Anamorphism by cementation tends to integrate rocks, in some cases with, in other cases without, the recrystallization of the original minerals. Anamorphism by rock flowage (dynamic metamorphism) tends to produce dense crystalline rocks, usually slaty, schistose, or gneissic, but sometimes lacking these textures, because of coarse recrystallization. At igneous contacts lamellar textures may be produced but not necessarily so; the rock may be rendered dense and crystalline without becoming schistose or gneissic.

In the development of slaty, schistose, and gneissose structures, the different minerals are more or less segregated in bands. Within these bands the minerals are of fairly uniform size, though not necessarily the size of other minerals in adjacent bands. The shapes of the minerals are those primarily determined by crystal habit. Measurements [1] show that on an average, the different dimensions do not differ greatly from the dimensional proportions of the same minerals in the igneous rocks, though on the whole they may be somewhat more elongate.[2] It frequently happens that in the later stage of schist development there are porphyritic growths which do not seem to be dimensionally controlled, or distorted or broken by movement, indicating their development after movement has ceased, though the pressure conditions may still be intense. (See p. 180.)

If Becke's term "crystalloblastic texture" be adopted for the schistose, slaty, and gneissic textures, it must be, according to the writers' view, with important qualifications. (See pp. 186–188.)

[1] Leith, C. K., Rock cleavage: Bull. 239, U. S. Geol. Survey, 1905, pp. 24–48.

[2] Trueman, J. D., The value of certain criteria for the determination of the origin of foliated crystalline rocks: Jour. Geol., vol. 20, 1912, pp. 236–241.
Leith, C. K., Rock cleavage: Bull. 239, U. S. Geol. Survey, 1905, pp. 24–48.

Not all gneisses are products of secondary anamorphism. Some are "primary" or formed before cooling of an igneous melt. Some are "injection" gneisses formed by the parallel injection of igneous material into laminated rocks.

MINERALS

The new minerals added during cementation may be the same as those already in the rock, as when quartz is added to quartz-sand, or they may be of a different type. The most abundant and permanent cement is quartz. Calcite, iron oxide, and various other minerals are locally important as cements. In rock flowage the essential change is in the direction of producing chlorite, hornblende, and mica, and other platy or columnar silicate minerals, from the carbonates, hydrates, and oxides in the sedimentary rocks, and from the destruction of the original feldspars and ferromagnesian minerals of the igneous rocks. Under certain conditions abundant secondary feldspars are formed. Rock flowage in extreme cases produces porphyritic garnet, staurolite, sodic and potassic feldspar, biotite, and other minerals, but usually in small amounts as compared with the minerals above named. Quartz may or may not be changed in amount during the process. Under contact metamorphism the same results may be obtained but at these places there may be additional mineral developments due to the introduction of materials from the magma.

COMPOSITION

During cementation, composition may be changed by the addition of substances from without where these are different from the substances already in the rock. Where cementation is accomplished by the addition of minerals like those already present, as cementation of quartz-sand by silica, the composition of the rock is not materially changed.

The anamorphism of sediments by rock flowage or contact action involves the expulsion of the constituents previously added from the atmosphere and hydrosphere, namely, water, carbon dioxide, and oxygen, the recombination of the alumina and silica with bases to produce silicates, and if our inferences on pages 202–203 are correct, the expulsion of any of the other substances

like quartz or lime, which may be present in the rock in excess of the requirements for producing the characteristic schist-making minerals. At igneous contacts, especially in limestone, there is some actual addition of substances from without. Dynamic metamorphism of igneous rocks involves hydration and other changes of a katamorphic nature, either during or prior to the deformation. That changes in chemical composition do take place during the processes of anamorphism is sufficiently well established on a quantitative basis to make chemical criteria based on the assumption of little or no change almost useless in the identification of schists and gneisses. (See pp. 226 et seq.)

VOLUME

Cementation of sediments involves change in volume, in so far as there is settling or slumping. On the other hand, the introduction of cements, which crystallize from solution, might have the effect in some cases of actually increasing volume, due to the crystallizing power of the cementing minerals. (See pp. 128–129.)

During anamorphism of sediments by rock flowage there is a diminution in volume due to reduction of pore space, due to the elimination of lighter constituents, like carbon dioxide, water, and oxygen, and possibly other substances, and due in some cases, to the development of minerals of higher density. It is true that the molecular volumes of certain minerals developed by anamorphism are greater than the molecular volumes of the replaced primary minerals. (See pp. 185–186.) For example, the molecular volume of wollastonite is greater than that of calcite, but wollastonite should be compared with calcite plus quartz, in which case a diminution in volume appears. Notwithstanding the fact that some of the minerals developed under anamorphism are of larger specific volume than the primary minerals from which they are derived, it is true that the characteristic schist-making minerals are, on the whole, of a higher specific gravity than those characteristic of katamorphism. In other words, proportions of hornblende, biotite, muscovite, and chlorite, common in schists, give distinctly higher specific gravity to the rock than a combination of kaolin, quartz, and calcite. When the decrease in porosity and the elimination of certain substances not needed for the schist-

making minerals are considered, a decrease in volume becomes self-evident.

Under contact metamorphism the volume may be likewise reduced, but where there is considerable addition of substances from the igneous rock, metasomatically replacing constituents of the sediment, there may be little or no change in volume.

ENERGY CHANGES IN ANAMORPHISM

Energy Changes of Chemical Reactions

Heats of formation of the mineral molecules involved in anamorphism are known in a number of cases, but sufficient data are not available to indicate closely what the energy change has been in any specific rock alteration. So far as the principal chemical changes in anamorphism are deoxidation, decarbonation, and dehydration, in other words, essentially the reverse of katamorphism, energy is absorbed. But there are a number of minerals characteristic of anamorphic rocks, such as sericite, chlorite, talc, zoisite, and epidote, which, because of their hydrous nature, suggest liberation of energy during their development. So far as they have been derived by partial dehydration of higher hydrates energy has been absorbed. If, however, they represent molecules formed during the earlier katamorphic phase of the cycle which have persisted and undergone recrystallization during anamorphism, their presence indicates neither absorption nor loss of energy during anamorphism. Grubenmann concludes, on grounds that do not seem entirely conclusive (pp. 188–189), that in an "upper zone" of schists, formed under moderate anamorphic conditions, energy is liberated by the chemical reactions, in a middle zone it is both liberated and absorbed, and in a lower zone, representing extreme conditions, it is absorbed. Van Hise [1] concludes that

"reactions take place with the liberation or absorption of heat, depending upon what is demanded by the pressure. Commonly the preponderant chemical reactions are those which take place with absorption of heat. . . . The rocks in this zone are under so great pressure in all directions that this fact demands chemical reactions which produce diminished volumes irrespective of whether heat is liberated or absorbed by them."

[1] Van Hise, C. R., A treatise on metamorphism: Mon. 47, U. S. Geol. Survey, 1904, pp. 167–168.

Our conclusion is, from the partial information available, that the chemical changes of metamorphism are dominantly endothermic and are, therefore, favored by increasing temperatures. This accords with the inference usually drawn from study of the location of anamorphic changes that high temperature is one of the important factors. The heat necessary for these changes may be inflow from without the system studied, or it may be developed within the system by the transformation of mechanical energy into this form.

Energy Factors in Volume Change in the Zone of Anamorphism

In general volume is decreased in anamorphism by (a) closing of voids, (b) chemical and mineral changes producing minerals of higher density, and eliminating certain constituents, principally water and carbon dioxide, (c) mechanical compression of the mineral constituents themselves. These factors involve absorption of energy by the anamorphosed rock by virtue of the transformation of mechanical energy into heat, and by virtue of storing of potential mechanical energy so far as minerals are deformed within their clastic limit.

Superimposed on these are the volume changes due to temperature variations. Increase in temperature tends toward increase in volume. If such expansion is prevented by pressure, heat is converted into potential mechanical energy in the compressed minerals. Pressure and heat act in opposite directions so far as volume changes are concerned and should be considered independently.

Decrease in volume is regarded by Van Hise as the controlling factor in anamorphism. In other words, the changes take place which the pressure requires, regardless of whether this means liberation or absorption of energy by the chemical reactions. The remarkable influence of crystal shape in determining the course of rock flowage (see pp. 204–205) likewise suggests control by physical conditions. Grubenmann argues that while the volume is decreased under moderate conditions of anamorphism, in extreme conditions (his "lower zone" of anamorphism) the volume may be increased. He bases his calculation, however, only on the comparison of the molecular volumes of the primary and secondary minerals, and does not take into account the changes in volume

by the reduction of pore space or the elimination of water, carbon dioxide, oxygen, or other substances, nor does he take into account the volume of the actual minerals destroyed.

Net Energy Change of Anamorphism

The net results of physical and chemical changes in anamorphism seem to be absorption of energy by (a) mechanical compression, (b) endothermic chemical reactions, which tend to transmute heat into the energy of molecular combinations, (c) increase in temperature, due to either mechanical changes or to inflow from outside sources.

DYNAMIC METAMORPHISM OF IGNEOUS ROCKS IN RELATION TO THE METAMORPHIC CYCLE

Where the typical products of dynamic metamorphism are produced from igneous rocks, there is question whether the change is really anamorphic. It is a fact that slightly weathered, and especially surface igneous rocks, are more likely to develop typical schists and gneisses than entirely fresh igneous rocks. In other words, anamorphism seems to act more effectively on rocks which have gone at least some distance through the katamorphic phase of the cycle prior to their anamorphism. In view of the fact that the typical schist-making minerals, like mica, chlorite, and hornblende, are slightly hydrous, there is a suggestion that their development is in the direction of katamorphism as compared with original igneous rocks, but whether the necessary water was introduced during or prior to the development of schists and gneisses is not clear. Whether the fresh igneous rocks can reach this metamorphic stage without going through any preliminary katamorphism we do not know. It has been suggested that under these conditions certain anhydrous gneisses may be formed by the granulation of the original constituents and development of secondary feldspar rather than schists characterized by the new development of hydrous schist-making minerals.

EXTENT TO WHICH ANAMORPHISM TENDS TO REPRODUCE IGNEOUS ROCK FROM SEDIMENTS

The anamorphic changes of sediments are in the general direction of producing rocks similar to igneous rocks, in that volume

and porosity are reduced, the oxygen, carbon dioxide, and water added during katamorphism are eliminated, and complex silicates are formed. In so far as materials are introduced by magmas along igneous contacts or the rocks are actually fused, they may take on the characteristics of igneous rocks. Feldspars may develop abundantly and the hornblende may go back to augite.

A comparison of Plates II, IV, V, VI, VIII, representing katamorphic alterations, with Plates X, XI, XII, XIV, XV, representing anamorphic alterations, indicates that anamorphism tends to reverse the process of katamorphism in the general direction of reproducing igneous composition. For instance, in the weathering of granite the principal end-product is clay from which silica has been taken away; lime has been eliminated more than magnesia, and soda more than potash. Carbon dioxide, oxygen, and water have been added. In Plate X representing the contact metamorphism of shale, it is clear that the reverse processes have been operating. Silica, as a whole, has relatively increased, lime increased in relation to magnesia, soda increased relative to potash, water, oxygen, and carbon dioxide have been eliminated. Naturally these changes are more marked in substances like limestone and quartzite, which depart most widely from igneous rocks in their chemical changes by katamorphism, and consequently show greater changes in going back toward the igneous type under anamorphism.

Yet it is equally clear that the schists and gneisses as a group lack certain constituents which are more abundantly present in the igneous rocks, namely, soda and lime, and that the schists usually contain minerals of different kinds and proportions than are characteristic of igneous rocks. The significance of these mineralogic differences are discussed below.

TENDENCY OF ANAMORPHISM BY ROCK FLOWAGE TO PRODUCE CERTAIN MINERALS DIFFERING FROM MINERALS OF IGNEOUS ROCKS

While anamorphism is clearly the reverse of katamorphism in its general tendencies, it is clear that the schists and gneisses resulting from rock flowage differ as a whole in certain important respects from igneous rocks, and that even igneous rocks under dynamic metamorphism may pass into schists and gneisses simi-

lar in all essential respects to those formed from the sediments. So much alike are the schistose products of sedimentary and igneous rocks that their origin constitutes a difficult problem. In other words, the tendency of anamorphism to produce rocks similar to igneous rocks seems to stop at a certain point where the characteristic schist-making minerals are formed, like mica, hornblende, and chlorite, together with quartz, and feldspar. If convergence toward these mineral types can be demonstrated, it may be used as a guiding principle in the study and interpretation of dynamic metamorphism.

Descriptions of anamorphic changes on the foregoing pages are sufficient to establish the fact that there is a marked tendency for development of sericite, mica, and hornblende schists from the dynamic anamorphism of several principal types of sedimentary rocks and from igneous rocks. A sericite schist may be developed from a quartzite, from an acid igneous rock, or, even in some cases, from a basalt. A hornblende schist may develop from a limestone, from a basic igneous rock, or in some cases, even from a granite or a shale.

The average igneous rock (p. 74) has about 50 per cent feldspar, 26 per cent ferromagnesian minerals, and 20 per cent quartz. The schists or gneisses developed from these igneous rocks have a considerably smaller proportion of feldspars. The ferromagnesian minerals, instead of consisting dominantly of augite and biotite, with some muscovite, are now largely hornblende and muscovite, with subordinate quantities of biotite.

Whether the parent rock is igneous or sedimentary, or whatever its mineral content, the resulting schists and gneisses are characterized by hornblende, chlorite and mica, in some cases feldspar, which are developed to such an extent that the very nature of the original rock is often lost. If the mineralogical change were not extensive the problem of origin of schists and gneisses would not be nearly so difficult as it is. The very existence of the problem testifies to the great mineralogical changes which have occurred.

A few distinctive minerals, such as mica, chlorite and hornblende, sometimes feldspar and other minerals, are the characteristic resulting products of anamorphism or dynamic metamorphism of a considerable variety of parent rocks. In this sense, then, there is a convergence towards a certain mineralogic type.

Chemical Changes Involved in Convergence to Schist-Making Minerals

When the chemical changes in anamorphism and rock flowage are considered, there is less general acceptance of the fact that actual changes in composition take place and that these changes are in the direction of producing the composition of the characteristic minerals in the end-products. Probably it is the general view that the conditions of rock flowage and anamorphism are not usually favorable to extensive transfers of materials involved in change of composition. It is often true that there are no important changes in percentages of chemical constituents during the process of rock flowage and anamorphism, the mineralogical change having been accomplished through recrystallization of substances present. This is especially true when the original rocks are well adapted in their composition, like shales, to the production of the necessary schist-making minerals. But even in these cases the elimination of water, carbon dioxide, and oxygen is recognized. We think it can be likewise inferred, in cases where the composition of the parent rock differs widely from the composition of the hornblende, mica, or chlorite, that important chemical changes take place, principally by the elimination of the substances present in excess of these requirements, but possibly in some cases by addition of substances from without, and that there is therefore a chemical, as well as mineralogical, convergence toward mineral types.

It is not easy to find in the field satisfactory gradations from massive to schistose rocks where one can be certain that the mass was originally of uniform composition throughout. It is still more rare that these gradations have been thoroughly sampled so that the analyses furnish an adequate basis for comparison. It is significant, however, that where this has been done (and we have searched the literature carefully) important changes are to be noted. (See Pls. XI, XIII.)

Fortunately we are not obliged in our conclusion to rely entirely on the few well sampled gradations from primary to schistose rock. A study of the actual compositions of the principal groups of schists brings out the fact that these vary in certain essential respects from those of the primary rocks from which

they are supposed to have been derived. In Figs. 14, 17, and 18 analyses of common schists have been platted graphically in order that they may be compared with both known and possible parent rocks. An examination of these plates brings out the fact that the composition of the schist tends to approach the distinctive chemical characteristics of the dominant platy or columnar mineral in the schist. This is especially well shown in the sericite schist, the composition of which lies between that of the primary rock and that of the mineral sericite. It is indicated not only by the position of the analyses on the diagram but by the shapes of the flags showing the relative amounts of constituents. If we were to include in these plates only those schists in which the processes of anamorphism and rock flowage have gone to an extreme, the tendency would be still more obvious, because rocks are included in these diagrams, described as hornblende, chlorite, or mica schist, which have but a small proportion of these minerals.

Our inference from the available facts is that, while recrystallization of substances present has of course played an important part in the production of schists, for some rocks important changes in composition have also occurred; that these changes have tended to give the rock a composition in which the influence of mica, hornblende, or chlorite is clearly discernible, that these changes are known both in sedimentary and igneous rocks, of both acid and basic composition, and that the changes have been sufficiently important to make it impossible, along with other reasons, to use chemical composition as a conclusive criterion for the identification of origin of schists and gneisses.

Significance of Convergence

If this idea of convergence be correct, our attention is directed to the physical and chemical characteristics of a few minerals like mica, chlorite and hornblende, as important factors in anamorphism by rock flowage. Obviously they are adapted to the conditions of rock flowage; otherwise, they would not develop at the expense of other minerals. It is not so clear whether they are adapted by their crystal habits, by their cleavages, by their composition, by the temperatures required for their development, or by a combination of these qualities.

According to Becke [1] these minerals, on the whole, require lower temperature for their development than the average minerals of the igneous rocks. He regards this as an important reason for their development and thus explains the difference in composition between the schists and the igneous rocks. Certain it is that they often develop under conditions of dynamic metamorphism far from the influence of igneous rocks and oftentimes not far from the surface. It is difficult to determine the temperature resulting from dynamic action, but it is easy to believe that it was lower on the whole than during the crystallization of magmas. Experimental work on the melting temperatures of minerals is not conclusive on this question for the reason that the conditions of experiment are clearly not the conditions of making of a crystalline schist. Clarke [2] suggests that temperature of crystallization should not be assumed to be the same as that required for the crystallization of the mineral from the melt.

Some few minerals like quartz have differing crystalline form above and below inversion points, and, therefore, may be used as geologic thermometers.[3] Inversion points have not been found for the characteristic platy minerals of schist. Evidence of this kind from the quartz in schists is very fragmentary and inconclusive, some of it showing by its crystal form that it has developed above the inversion point of 575° and some below this temperature. The part temperature may have played in determining convergence toward the few characteristic mineral types in schists is yet to be proved.

Another characteristic of the prominent minerals of schists seems to the writers to play a considerable part in their development. The fact that they are always arranged according to their dimensions, their greatest mean and least being respectively parallel, and the fact that their dimensions in the schist are clearly those determined by their habit of crystal growth, rather than by any growth or breaking giving other shapes, direct our attention to the crystal habit of these minerals as essential reasons for

[1] Becke, F., Über Mineralbestand und Struktur der kristallinischen Schiefer: Compt. Rend. IX Cong. geol. internat., Vienna, 1903, pp. 553 et seq.

[2] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, p. 281.

[3] Wright, F. E., and Larsen, E. S., Quartz as a geologic thermometer: Am. Jour. Sci., vol. 27, 4th ser., 1909, pp. 421–447.

their development. Whether processes of rock flowage have been weak or intense, the resulting minerals maintain their habit and dimensional characteristics. It seems to be something more than a coincidence that "crystallizing force" has been able to exert itself to the extent of causing important mineralogical and chemical changes toward a limited number of mineral forms, under such a variety of conditions and with such variety of available materials. It has drawn together the materials needed, eliminated those not needed, and had developed crystals of uniform habit. Under given environment it has had the capacity to organize the substances in a fashion best adapted to environment in much the same way that organisms have been supposed to adapt themselves to environment. The problem before us may be similar to the biologic question as to whether life forms are distinctly the results of physical and chemical environment or whether that mysterious force we call life on occasion rises superior to environment and to some extent modifies and controls external conditions. It is sometimes said that man is the multiplier and environment the multiplicand in the product determining history. Following the simile, it may be that the forces of crystallization are the multiplier, environment and rock composition the multiplicand, and the schists and gneisses the product.

Schists, slates and secondary gneisses form a group which may be regarded essentially as a by-product of the metamorphic cycle, in that the processes involved have failed to reproduce primary igneous rocks, from which all metamorphism presumably starts. Fusion along igneous contacts and introduction of magmatic materials may close the gap between this group and the igneous rocks. Whether the metamorphic cycle has been completed in this manner for any great mass of metamorphic products is a doubtful question which is discussed in another place (pp. 268 et seq.).

PLACES AND CONDITIONS OF ANAMORPHISM

In the foregoing discussion emphasis has been placed on inductive study of anamorphic changes in rocks with only incidental reference to the places, conditions and agents favoring anamorphism. It is our view that this method of approach, being based more specifically and directly on facts of observation, is of more

service to the student in his field and laboratory work than a discussion based primarily on loci and conditions of metamorphism. The physical chemists and the geophysicists are supplying much valuable data on the conditions essential to alteration, but it is scarcely possible yet to indicate the precise nature of the control of metamorphic alterations, even in the simplest cases. It is not our purpose to attempt anything more than the briefest discussion of places and conditions of anamorphism, with a view of indicating some of the more important conclusions that have been reached.

Places and Conditions of Cementation

The early stages of anamorphism, here included under the general term *cementation* begin near the surface and extend to almost any depth. They merge by imperceptible stages into dynamic and contact metamorphism. Cementation may begin when sediments are first deposited either on land or under water, and may be far advanced before subaqueous sediments emerge from the water. Cementation appears most conspicuously immediately below the water table and downward to the bottom of the zone of rock fracture. This has been called the *belt of cementation* by Van Hise.[1] It does not follow that this is the only place where cementation occurs, or that all rocks within this belt are undergoing cementation, for certain rocks, such as the igneous rocks, may be breaking down by katamorphism, while others are being built up by cementation, and a certain amount of cementation may occur quite to the erosion surface.

Cementation is accomplished under a great variety of agencies and conditions. Cementation by circulating aqueous solutions carrying cementing materials is important. Another important factor is sufficient pressure to squeeze out the water and close up the openings. The pressure is mainly gravitational, but tangential stresses may be locally effective. So far as the pressure brings about interior movement in the rock mass, it favors development of parallel arrangement of mineral constituents, but in this stage movement seems to be so slight that results of this kind are insignificant. Recrystallization of minerals within the rock, some-

[1] Van Hise, C. R., A treatise on metamorphism: Mon. 47, U. S. Geol. Survey, 1904, pp. 162–163.

times leading to a marked change in volume, accompanies cementation. The selective absorption of materials from the solutions by the minerals in the rock is a factor. High temperature does not seem to be an essential requirement for the reason that cementation seems to occur equally well at the surface and to considerable depths, near igneous rocks and far from them.

Dynamic Metamorphism or Rock Flowage

Conditions. Dynamic metamorphism or rock flowage requires surrounding pressures sufficient to prevent the existence of openings, with differential pressures capable of causing movement in the rock. Whatever the pressures, dynamic metamorphism does not result unless there is rock movement. The extent and degree of alteration appear to depend on the ease with which the rock yields mechanically to these pressures, ease of recrystallization of its mineral particles, the kinds of minerals present and their compositions, the temperatures, the nature and amount of the solutions, and the extent to which they have access to the rock, etc., all of which factors are more or less interrelated. Anamorphic changes by dynamic metamorphism seem to be favored by conditions of some depth. Depth, however, is not alone sufficient. There are cases of deep burial of sediments in which anamorphism has progressed to a slight degree. On the other hand, schists and gneisses have been produced very near the surface. One of the important factors is the competence of the rocks to resist rock flowage. A weak rock is often recrystallized while a strong rock is comparatively little affected. It is obvious that a soft shale or limestone may yield and become anamorphosed under dynamic stresses relatively near the surface, while a hard quartzite or igneous rock may be unaffected to a considerable depth.

Ordinarily it is supposed that temperature developed by movement may be an important factor in anamorphism, but this is difficult to prove. Becke [1] has pointed out that the minerals in schists seem to require lower temperature for their development than the minerals in igneous rocks, and yet even this conclusion is doubtful. (See p. 183.) So far as can be told from the nature of the minerals present, the changes may occur under a very

[1] Becke, F., Über Mineralbestand und Struktur der kristallinischen Schiefer: Compt. Rend. IX Cong. geol. internat., Vienna, 1903, pp. 553 et seq.

wide range of temperature, so wide a range, in fact, that one is compelled to conclude that conditions other than temperature may control. On a preceding page we have discussed the probability that the crystallizing power of certain minerals may be an important factor in determining what minerals shall develop and consequently what changes in composition will ensue during dynamic metamorphism. This seems to be at least one factor common to a wide variety of conditions, both of materials and environment, in the production of crystalline schists.

Places. For the above reasons, rock flowage may occur in incompetent rocks or in strong rocks locally yielding to great stress differences, or in rocks of a composition susceptible to change, or in rocks capable of yielding certain metamorphic minerals best adapted by habit to these conditions, or in rocks affected by igneous intrusion, while near at hand, at the same depths, rocks not so situated, either by reason of their composition, competency, or the conditions effecting them, may not undergo anamorphism, and may, in fact, be undergoing katamorphism. In this sense the zone of anamorphism must be regarded not as a zone of uniform depth, but rather as a more irregular zone determined by a variety of conditions above given.

Recrystallization. The process by which the minerals of a schist and gneiss are developed and arranged may be conveniently designated by the general term "recrystallization." The precise nature of this process is not yet clear. It involves molecular rearrangement and presumably movement of materials by other than mechanical means. Recrystallization seems to be favored by the presence of water, as indicated by the ease of recrystallization in rocks carrying water and the development of hydrous platy and columnar minerals. Rocks containing little or no water may, however, be thoroughly recrystallized under dynamic metamorphism, as for instance, the carbonates. Molecular rearrangement and recrystallization without the aid of a fluid solvent is illustrated in the crystallization of cold metals and the formation of alloys by solid solution. Experimentally, it has been shown that recrystallization may occur under dry conditions, as in the heating of certain powders below the fusion point.

Comparison of analyses of rocks in successive stages of dynamic anamorphism shows clearly defined elimination of water and hence

notable movement of water during anamorphism. The fact that openings tend to be closed by dynamic metamorphism makes it difficult to understand how water can move through these rocks with any degree of freedom. It is necessary to assume that it is present in capillary and subcapillary openings. Since the movement of water is imposed by pressure conditions, it is probably resisted rather than aided by phenomena of adhesion and capillarity. Capillary forces are effective only when there is a free liquid surface within the pores; moreover, they diminish steadily with rise of temperature and vanish at the critical point of the liquid. Calculations show that the effects producible by capillary attraction at any considerable depth are insignificant in comparison with the pressure due to the hydrostatic column, except in pores of such fineness that the amount of water which would flow through them is infinitesimal.

It is highly probable that the only source of water as an agent in recrystallization is the free or combined water of the rock itself and that there is no penetration of water from outside sources during anamorphism. From the result of certain experimental work, following the lines of Daubree's experiment, on the passage of water through a disk of sandstone against a certain counter pressure of steam, Johnston and Adams [1] conclude:

"it appears therefore as if the probabilities were all against the notion that appreciable amounts of meteoric water can ever penetrate into deep-seated and highly heated rock masses." There is "difficulty of imagining a reasonable configuration of the rock at depth, such as would insure that the total pressure within the pore is not overbalanced by the pressure to which the plastic rock surrounding it exerts."

Starting with the assumption that the crystal is in contact with a solution, it is conceived that when under pressure the solubility of the strained face is increased and material is removed from this place to be redeposited where the strain is less, tending to make the particle ellipsoidal or elongate.[2] This conception meets difficulties when we remember that the shapes of the mineral

[1] Johnston, John, and Adams, L. H., Observations on the Daubree experiment and capillarity in relation to certain geological speculations: Jour. Geol., vol. 22, 1914, p. 15.

[2] Johnston, John, and Niggli, Paul, The general principles underlying metamorphic processes: Jour. Geol., vol. 21, 1913, p. 610.

particles in schists seem to be determined almost solely by crystal habit. No matter what the flattening or degree of movement of the rock mass, the shapes of the individual crystals remain nearly the same. If the crystal has grown in length, it has likewise grown proportionately in thickness. In other words, it has added material on the strained as well as the unstrained faces. Observation shows that the minerals originally present diminish in proportion during the progress of dynamic metamorphism, their places being taken by new minerals, adapted by habit to the new conditions. In other words, the materials which go into solution at points of strain seem to go to the building up of flat or elongate crystals of a new generation, rather than to the building up of unstrained ends of crystals not fortunately oriented or adapted by their shape to these conditions. Where not properly oriented the minerals do not grow.

These considerations suggest a modification of the theory that recrystallization takes place by solution of material on strained faces with addition to the unstrained ends, for the evidence clearly shows that material is added both on the ends and on the strained faces opposed to the pressure in order to maintain the crystal habit. The flattening of the rock as a whole is accomplished by the redistribution of substances, partly by granulation and partly by growth of new minerals of flat or elongate habit better adapted to the new conditions. Recrystallization dominates over granulation in the later stages of the process. Flattening of the rock mass as a whole in one direction does not mean that solution is taking place on all crystal faces opposed to this direction. It is taking place on part of them, on others matter is being added. While there is net loss of material in the direction of greatest shortening, this is accomplished despite the fact that new individual crystals actually grow in this direction. In homogeneous material of low viscosity the pressure conditions will develop highly elongate and flattened mass, but where the units are crystals the flattening may go only to the extent which is allowed by the crystal habit. The crystal habit in other words interposes a force of its own to prevent the fullest compliance with the differential movement. This conception of recrystallization puts the emphasis on movement rather than on differential pressures in forming schistose textures. Under the usual theory of recrystallization, differen-

tial pressures insufficient to cause movement would still cause strain on faces opposed to the greatest stress, and therefore tend to cause a parallel arrangement. Geologic and petrographic evidence, together with observations of the kind above stated, seem to us to put the emphasis on the *yielding* to these stresses, rather than on the differential stresses. The question is not the manner in which minerals resist differential stresses, but rather the manner in which they yield to these stresses.

Under conditions of high pressure and temperature in the absence of *differential* stresses sufficient to cause movement, certain porphyritic minerals of more or less equidimensional crystal habit develop (p. 180). The high density of these minerals suggests that they have formed in response to demand for less volume.

The growth of a crystal under pressure has been demonstrated experimentally by Becker and Day,[1] who show that the crystal growing from a solution exerts a pressure approximately as great as its crushing strength. When the least of the three principal stresses is in itself great, the mineral particles may take on a rigidity much greater than that ordinarily figured as their crushing strength. (See p. 179.) Possibly the crystal exerts much greater pressures than the crushing strength under these conditions, for they certainly grow under conditions of great compression in all directions and maintain their habit, but this question is complicated by the fact that the growing crystal does not actually force the rock mass apart in all cases, for it may make room for itself by absorption and destruction of adjacent crystals.

Contact Metamorphism

It is obvious that igneous intrusion develops conditions of high temperature and pressure favorable to anamorphism, and tends greatly to accelerate the process. It is almost impossible in some cases to discriminate between purely dynamic metamorphism and that due to igneous intrusion except where there is absence of schistose structure or good evidence of actual transfer of materials of the igneous rocks into the metamorphic rocks.

Conditions at some igneous contacts indicate clearly that both

[1] Becker, G. F., and Day, A. L., The linear force of growing crystals: Proc. Wash. Acad. Sci., vol. 7, 1905, pp. 284–287.

temperature and pressure are important. Temperature is clearly more important than in dynamic metamorphism. Also, it is more evident that aqueous and gaseous solutions permeate the rock. This has been directly observed in the vicinity of volcanoes, and is definitely inferred from the fact that there have been actual additions from the magmatic sources to the rocks under such conditions. The precise range of control of temperature, pressure, and the nature of the solutions, is not known. Experimental work on the metamorphism and synthesis of a few minerals makes certain inferences possible, but even in the simplest cases it is difficult to measure and control all of the considerable number of variables that enter into the experiment. Owing to the lack of requisite quantitative data, no statement as to the effects produced in a given system by change of any of the factors can yet be made. Discussing the application of certain physical-chemical laws to metamorphism, Johnson and Niggli,[1] of the Geophysical Laboratory of Washington, conclude that

"the general application of experimental results which obtain for a given system under given external conditions to another system under similar conditions, or even to the same system under widely differing conditions, is subject to considerable limitation. Conclusions drawn from such extrapolation of experimental evidence will commonly be of little value, and may be altogether misleading; moreover, one may as well guess the final result as arbitrarily choose the data required in calculating it."

[1] Johnston, John, and Niggli, Paul, The general principles underlying metamorphic processes, Jour. Geol., vol. 21,1913, p. 623.

PART III

GENERAL DISCUSSION OF METAMORPHISM

CHAPTER I

DETERMINATION OF ORIGIN OF METAMORPHIC ROCKS

In the foregoing chapters we have attempted to follow the known changes of sedimentary and igneous rocks into their metamorphic equivalents. In field and laboratory study of metamorphic rocks it is usually necessary to work backward in ascertaining their origin, which requires a somewhat different point of view. The purpose in the present chapter is to discuss field and laboratory methods of determining origin of these rocks. The principles discussed in preceding chapters are of course important considerations in this type of work. Ordinarily the discussion of the origin of metamorphic rocks is confined to the schists and gneisses. In a broader sense the sediments and residual deposits are metamorphic products, the origin of which it is essential to determine, especially in field studies of deeply weathered areas and in investigations of problems of sedimentation. The problem of the origin of these materials is similar in its essential respects to that of the origin of schists and gneisses. In both, metamorphic criteria play a large part; in both other criteria are important. To go thoroughly into the criteria of origin of sedimentary and residual deposits would involve the consideration of a wide range of conditions entering into sedimentation, which is beyond the scope of this book. The inductive development of this subject from facts of observation has not been satisfactorily accomplished. In the following paragraphs a few of the more obvious criteria of origin of sediments and residual rocks are mentioned with a view of emphasizing the fact that metamorphic principles enter into their study quite as much as into the study of the schists and gneisses.

ORIGIN OF RESIDUAL CLAYS AND SOILS

In regions where the residual accumulations have not been disturbed or distributed by transportation, it is often possible to trace

direct gradation of clays and soils to the parent rock. This is purely a field problem. Where erosion has not exposed the fresh rock it is often necessary to infer its character from the residual material. A clay soil containing chert fragments may in many cases be referred to the weathering of a limestone or dolomite formation. A sandy soil may be referred to sandstone, or in some cases, to acid igneous rock. A black residual soil, without quartz, may be referred to the weathering of basic igneous rocks. Not infrequently, a soil exceptionally high in iron may be referred to some basic type. In some cases weathering produces an iron ore of the Cuban lateritic variety. Bauxitic alteration usually refers to some type of acid igneous rock, although it may include basic igneous rocks or sediments as well. The texture [1] of the weathered residual may often be that of the rock from which it is developed, particularly where developed from coarse-grained igneous rocks, although in general, as the surface is approached, there is a tendency for original textures to be destroyed and in some cases to be replaced to some extent by granular texture. The color of the soil may often give a clue as to what is underneath. This is, however, so dependent upon climatic conditions and on minor constituents of the rock that color criteria must be worked out separately for each locality. Within a given district differences in color often serve as very satisfactory guides to the determination of the source of the weathered material. A criterion often used in tropical countries is the washing of the soils for the purpose of separating the heavy minerals.[2] These are persistent through weathering and may indicate the type of rock from which the soil is derived. For instance, if a soil is found to contain zircon, apatite, garnet, or monazite, with well developed crystal surfaces, it is a fair presumption that it is residual from igneous rocks. If these same minerals are rounded, it very likely indicates sedimentary origin. The proportions of these substances in the residual mantle sometimes distinguish between several igneous rocks of a region.

The literature on soils contains many references to criteria of the origin of soils which vary so much with different climatic and

[1] This term is used in the petrographic sense, and not, as used by the soil physicist, to indicate specifically size of ultimate particles.

[2] Derby, O. A., On the separation and study of the heavy accessories of rocks: Proc. Rochester Acad. Sci., vol. 1, 1891, p. 198.

topographic conditions and with minor constituents of the original rocks that it is difficult to frame any general statement. Systematization of this subject offers an attractive field for investigation.

DEPOSITS OF GLACIAL ORIGIN

Products of glacial origin may be identified as such by the forms of the deposits. Their texture gives evidence of their origin, particularly their content of fragments of widely varying size, as in the boulder clays. There may be some lack of correspondence of glacial products with the immediately underlying rocks, though near-lying rocks may constitute a conspicuous part of the glacial materials. Chemical and mineralogical composition serve as a guide, for the glacial products consist of ground up rock material from which the bases have often not been removed by solution to the same extent as in weathering. The striation of pebbles is often a decisive criterion. In well-weathered till of early glaciation these criteria fail.

Water deposited glacial sediments may be identified by their stratigraphic and petrographic relations to other deposits of glacial origin and by their lack of weathered materials, but where these are carried far from their source, as on the Alaskan coast where they include marine fossils, their origin becomes obscured.

TRANSPORTED CLAYS AND MUDS

During transportation there has been more or less sorting, resulting particularly in the separation of sand and the deposition of muds in beds of widely varying texture and composition. This process of sorting tends to obscure evidences of origin. It makes it practically impossible to trace any given mud to its source on the basis of the characteristics of the mud itself. Where the transportation channels are known the contributing areas may be identified. Clays and muds may come from almost any type of rock except pure quartzite and pure limestone. Usually the limestones have enough impurities to yield, on weathering, considerable amounts of clay or mud. All igneous rocks will yield some of this material. A highly siliceous clay suggests derivation from acid igneous rocks or from siliceous sediments. A clay lacking quartz suggests a more basic source. A clay containing a

large percentage of bases suggests that it has been derived from glacial deposits or other material which has not undergone leaching, or that it may have been derived from volcanic muds as is suggested by the not uncommon association of mud deposits with basic igneous flows. The determination of the composition of the muds of the Upper Huronian of the Lake Superior region,[1] which are associated with tuffs and volcanic flows, shows that the muds are much nearer in composition to the basic igneous rocks and to the hydrothermal alterations of basic rocks than they are to the normal types of residual soils or average shales or muds.

Clay and mud are nearer the composition of average igneous and crystalline rocks than any other types of sediment. In Chapter VII, Part I, it is shown that the argillaceous sediments greatly predominate over the sands and carbonates in the products of igneous rocks and that the carbonates and sand may be regarded merely as by-products in the development of mud. The nearness in composition of the muds to average igneous and crystalline rocks makes it possible to infer a wide variety of sources, which makes the problem of identification of specific source a difficult one.

Heavy residuals are not useful as a criterion for determination of the origin of the muds, as they are in the sands, for the reason that they are ordinarily deposited with the sands and not with the muds, presumably because of their size and weight.

It is usually desirable to ascertain not only the origin of a clay or mud, but the physiographic and erosional conditions at the time of its deposition. The thickest mud and shale deposits seem to be those of deltas. The rapid alternation of beds, cross bedding, ripple and stream marks, and layers high in organic material are usually characteristic of such deposits. If the wave and current action is strong the muds may be swept for many miles along the coast. Ordinarily they are deposited farther from shore than the sands in such cases, but this is not always true, for there are many places where, because of peculiarities of current or shore line, the mud constitutes the shore material. Mud deposits are often interpreted by physiographers as representing shore deposits due to deep weathering followed by erosion, and these conditions are ordinarily correlated with low relief. These are merely

[1] Van Hise, C. R., and Leith, C. K., The geology of the Lake Superior region: Mon. 52, U. S. Geol. Survey, 1911, pp. 610–614.

suggestions of some of the inferences to be drawn from transported clays and muds. The list is incomplete; in fact, the subject remains to be organized in a comprehensive fashion.

SANDS AND SANDSTONE

Sands may be derived from almost any rock except the basic igneous rocks and pure limestones. Its ultimate source is very largely in the granites or other acid igneous rocks. In sandy deposits of arid regions, both subaqueous and subaërial, lack of decomposition and the high content of original silicates may lead to identification of original source. Perhaps the best criterion for tracing origin of the sands and sandstones is the presence in some of the sands and sandstones of certain accessory minerals of the zircon type which may suggest the nature of the source. For instance, the monazite sands on the east coast of Brazil suggest derivation from certain granitic rocks which have been found to contain monazite abundantly. It is sometimes possible to identify quartz grains from igneous rocks by the nature of the inclusions.

The physiographic and erosional conditions of derivation and deposition of sand may be inferred only very vaguely with the present state of knowledge. The sand may be regarded as essentially a by-product of the deposition of muds. It is ordinarily regarded as a shore deposit where there has been considerable sorting action of the waves. Dredging operations, however, sometimes show sands lying in patches in the muds and clays or lying off shore from the shales. There are many thick beds of quartzite (up to several thousand feet) the conditions of deposition of which cannot be satisfactorily inferred.

CALCAREOUS SEDIMENTS

Calcareous sediments may be derived from preëxisting limestones or from basic igneous rocks. Granites and quartzites are deficient in lime necessary for the production of limestone. The source of some limestone now being deposited can be traced by transportation channels, as, for instance, the calcareous deposits of the Gulf of Mexico are fed in large part by lime carried by the Mississippi River and obviously derived principally from Paleo-

zoic limestone formations. It is impossible except in rare instances to apply this method to older limestone formations, and no other way has yet been developed for tracing their origin. Limestone deposition has ordinarily been assumed to indicate land conditions of rather complete weathering as opposed to dominant disintegration. It has been commonly assumed that limestone indicates deep off shore conditions of deposition; however, most of the limestones over the continents have probably been deposited in shallow waters within range of the sunlight believed to be necessary for many of the lime depositing organisms.[1] The present distribution of limestone deposition is very erratic. In some places it is being deposited close to shore, in others off shore, in some cases in shallow water, in others in fairly deep water. The factors in the problem are ocean currents, temperature, distribution of light, precipitating agents, the mixing of river with ocean water, etc.

CRITERIA FOR DISTINGUISHING PRIMARY GNEISSES FROM METAMORPHIC GNEISSES

The term "primary gneisses" is used for certain banded igneous rocks whose foliation was produced during the consolidation of the rock. They are not the result of anamorphic processes acting upon crystallized rocks. There is much difference of opinion about the abundance of primary gneisses, due largely to the difficulty of distinguishing such gneisses from gneisses which are formed by anamorphic processes. Some of the criteria for distinguishing the primary gneisses are summarized by Trueman[2] as follows:—

"*Field evidence:* Banding in apophyses from the gneiss parallel to the walls and at an angle to the schistosity of the enclosing rock,[3] dikes of pegmatite belonging to the same magmatic series as the gneiss and either parallel to the gneissic structure and foliated with it or cutting the gneissic structure and undisturbed; lack of sharp contact between the acidic and more basic portions of the gneiss, indicating high temperature during the

[1] Schuchert, Charles, Paleogeography of North America: Bull. Geol. Soc. Am., vol. 20, 1910, pp. 438–439.

[2] Trueman, J. D., The value of certain criteria for the determination of the origin of foliated crystalline rocks: Jour. Geol., vol. 20, 1912, p. 231.

[3] Gregory, J. W., Quart. Jour. Geol. Soc. London, 1894, p. 265.

solidifications of the different bands;[1] presence of inclusions of foreign rock, which are but slightly deformed, in a matrix of well banded gneiss;[2] presence of distinct bands of widely differing composition, none of which may show evidence of shearing; flowlike curves of the banding, some of which may close in a circle."

"*Mineralogical evidence:* Presence of minerals formed characteristically only from igneous melts and arranged in a manner impossible of formation from solid rocks by metamorphism, e. g., nepheline and olivine; textures due to crystallization from an igneous melt. Weinschenk,[3] considers that epidote, garnet clinozoicite, sillimanite, and chlorite crystallize from the magma in the case of primary gneisses on account of the pressure present during solidification of the rock, but the exact state of the rock during their formation is not definitely known."

Becke emphasizes texture as a criterion for separation of the primary and metamorphic gneisses. (See pp. 186–187.) He regards the forms of the minerals in igneous rocks as dependent largely on their order of crystallization, that is, the earlier formed minerals have good crystal outlines, while those of later development, having been compelled to occupy the remaining spaces, are irregular in form. In the metamorphic gneisses, on the other hand, his view is that the crystallization of all the minerals has been more or less hindered by the solid condition of the rock and that the minerals possess the best developed forms which have the strongest crystallizing force, these being minerals which, on the whole, have the smallest molecular volume or the highest specific gravity. To the secondary texture, developed by rock flowage, Becke[4] gives the name "crystalloblastic texture." The characteristics of this texture are summarized on pp. 186–187.

Trueman points out that some of the features of mineral elongation ascribed as characteristic of the crystalloblastic structure of metamorphic gneisses are, as a matter of fact, possessed by the primary gneisses, and that the texture of the primary gneisses as a whole appears to be intermediate between the igneous and metamorphic types, being more like the latter, according as the

[1] Geol. Survey Can., Mem. 6, 1910, p. 83.

[2] Geol. Mag., new ser., decade 4, vol. IV, 1897, p. 354.

[3] Weinschenk, E., Congres' geol. inter., compte rendu, session VIII, I, 1900, p. 340.

[4] Becke, F., Über Mineralbestand und Struktur der kristallinischen Schiefer: Compt. Rend. IX Cong. geol. internat., Vienna, 1903, pp. 553 et seq.

movement producing the banding continued late during the period of the consolidation of the rock from the melt.

The use of texture as a criterion for the identification of primary gneisses seems on the whole, then, to be only of limited application, even if it be true that the primary gneisses as a group have textures of igneous rocks while the secondary gneisses have, as a whole, the crystalloblastic textures.

The problem of distinguishing origin of gneisses is even more complicated for the so-called injection gneisses, formed by the intrusion of igneous material in thin leaves between laminæ of the country rock. These gneisses may have characteristics both of primary and metamorphic gneisses.

IDENTIFICATION OF METAMORPHIC SCHISTS AND GNEISSES

Not only does rock flowage tend to obliterate primary textures but it modifies the chemical and mineralogical composition. In proportion, then, as rocks have undergone rock flowage, there may be difficulty in ascertaining their origin.

Field Relation as a Means of Identifying Schists and Gneisses of Metamorphic Origin

Field and microscopic observation of gradations from undeformed rocks into schists (or slates) or gneisses gives certain empirical methods for recognition of origin. For instance, a shale alters to a slate and this in turn to a phyllite. While it is difficult from the study of the phyllite alone to determine its origin, it so often has been observed as the end-product of this series of changes that there is little danger of mistake if it is referred back to a shale or a mud. A sandstone or quartzite may be traced into a mica-quartz-schist, seldom into a hornblende schist. A similar schist may be derived from the secondary deformation of certain acid igneous rocks. A quartz-mica-schist therefore is regarded as the natural development of an acid rock, but whether sedimentary or igneous may be doubtful, when field relations do not decide. A basalt is observed to grade into a chloritic and micaceous schist. The same result may be observed where certain shales are altered. Basic igneous rock (especially in the vicinity of intrusives) by rock flowage may pass into coarsely crystalline hornblende schists or gneisses. Amphibolites are known to be formed also by altera-

tion of limestone. Some banded gneisses, by their association with, and gradation to, granites, and by their mineralogical composition, seem to be surely the result of rock flowage of a granite, though cases of proved gradation are rare. It has been observed, however, that certain sediments, such as an impure quartz sand, have gone over to gneisses with general aspects similar to those presumably developed from a granite. The passage of a dolomite into a talc schist is not uncommon.

Schists or gneisses may be interbedded with sediments and be themselves in beds strongly suggestive of sedimentary origin. They may be in an igneous complex and have irregularity of form or distribution or relations to adjacent rocks more characteristic of an igneous mass than of sedimentary beds. Some gneisses of the Laurentian, for instance, are clearly original igneous rocks intrusive into adjacent rocks. Where the schists or gneiss shows marked differences in composition in different beds or bands, and this composition is persistent throughout these bands for long distances, it is suggestive of sedimentary origin, especially if some of the beds have mineral or chemical composition of sediments. The Baltimore and Carolina gneisses of the Piedmont Plateau [1] and the Idaho Springs formation of the Georgetown area of Colorado [2] are of this type. Yet analagous structure has been produced, perhaps on a smaller scale, by injections of igneous masses along parallel planes.

On the whole, with our present knowledge, field observations are likely to yield more satisfactory conclusions as to origin than other criteria below discussed.

Textures and Structures as a Means of Identifying Schists and Gneisses

Stratification or bedding, fragmental relicts, ripple marks, conglomerates, delta formations and abrupt changes in mineralogical composition and texture within very narrow limits are

[1] See Keith, Arthur, Washington folio (No. 70), Geol. Atlas U. S., U. S. Geol. Survey, 1900.

Mathews, E. B., Correlation of Maryland and Pennsylvania Piedmont formations: Bull. Geol. Soc. Am., vol. 16, 1905, pp. 329–346.

Bascom, F., Piedmont district of Pennsylvania: Bull. Geol. Soc. Am., vol. 16, 1905, pp. 289–328.

[2] Spurr, J. E., and Garrey, G. H., Economic geology of the Georgetown quadrangle, Colorado: Prof. Paper No. 63, U. S. Geol. Survey, 1908, p. 44.

evidences of sedimentary origin. When any of these are distinctly preserved the evidence is fairly decisive. There are many schists and gneisses, however, in which such evidences have been obscured by rock flowage and anamorphism.

The preservation of original forms of occurrence such as dikes, sills, and laccoliths, and preservation of distinctive igneous textures are evidences of igneous origin. In extreme cases of anamorphism and rock flowage these, however, are likely to be destroyed.

Mineral Composition as a Means of Identifying Schists and Gneisses

A great preponderance of quartz is perhaps more often characteristic of a sedimentary than an igneous rock. Where a gneiss or schist is dominantly quartz, one looks for other evidences of sedimentary origin. But the existence of highly quartzose rocks of the pegmatite and alaskite types makes quartz content alone a doubtful criterion. Preponderance of calcite is more satisfactory evidence of sedimentary origin.

Quartz crystallizes in one form above 575° C., in another below 575°. When, therefore, quartz is found in a gneiss to have crystallized below 575° this may be suggestive that the temperatures have not been those of fusion but that the quartz is the result of recrystallization of a rock at lower temperatures. This favors sedimentary origin. It is possible, of course, that quartz, originally crystallized above 575°, might recrystallize at lower temperatures and take on another form.

The abundant development of aluminum silicate minerals such as staurolite and sillimanite [1] has been more commonly observed in metamorphosed sediments than in igneous rocks. Any of these minerals, however, may be found also in igneous rocks.

Where gneiss is strongly feldspathic it is not likely to be regarded as of sedimentary origin. Yet so far as the sediment is undecomposed, it may be largely feldspathic, and also the anamorphism of a non-feldspathic sediment might make it feldspathic, though it is a question whether to a degree common to many gneisses.

The presence of graphite disseminated evenly through a band

[1] Emmons, W. H., and Laney, F. B., Preliminary report on the mineral deposits of Ducktown, Tennessee: Bull. 470, U. S. Geol. Survey, 1911, p. 158.
Spurr, J. E., and Garrey, G. H., Economic geology of the Georgetown quadrangle, Colorado: Prof. Paper No. 63, U. S. Geol. Survey, 1908, p. 44.

or zone becomes presumptive evidence of sedimentary origin, especially where, as in the Adirondack graphites, there are other evidences present.[1] Some graphite may be igneous in origin, but when evenly distributed in amount up to about 6 per cent in a generally slaty or quartzose zone, the hypothesis of igneous origin becomes untenable.

Mica or chlorite or hornblende affords no satisfactory criterion of identification of origin, for these minerals develop both from sedimentary and from igneous rocks. But so far as present evidence goes, they seem to develop more readily from sediments than from igneous rocks, perhaps because water is necessary. This criterion must be most carefully used, in view of the fact that sedimentary composition may be approached by the weathering of igneous rocks prior to anamorphism. The basalts of the Menominee district of Michigan [2] alter by katamorphism into chloritic rocks and under pressure alter to chlorite schists. The mineral change from the fresh rock is the same in both cases. It may be that the chlorite-schist stage was preceded by katamorphism of the basalt.

The separation of minute accessory constituents by washing is a means for identifying origin of schists and gneisses which has not yet been sufficiently used. Minerals of igneous rocks, like monazite, zircon, sphene, and garnet, are remarkably resistant to weathering, and remain in well-defined crystals in the residual mantle when all the other constituents have altered. When transported they become worn and rounded and tend to segregate with sand rather than the argillaceous sediments. In the schistose equivalent of the sand deposits the rounded grains persist, particularly the zircons, and afford evidence of sedimentary origin. Where in an argillaceous schist these heavy accessory minerals are lacking, a sedimentary origin is also probable. Trueman [3] has used this criterion with some success in several obscure cases.

[1] Bastin, E. S., Origin of certain Adirondack graphite deposits: Econ. Geol., vol. 5, 1910, pp. 134–157.

[2] Williams, G. H., The greenstone schist areas of the Menominee and Marquette regions of Michigan; a contribution to the subject of dynamic metamorphism in eruptive rocks: Bull. 62, U. S. Geol. Survey, 1890.

[3] Trueman, J. D., The value of certain criteria for the determination of the origin of foliated crystalline rocks: Jour. Geol., vol. 20, 1912, pp. 228–258; 300–315.

Identification of Schists by Chemical Criteria

The use of chemical criteria in determining the origin of schists (and some gneisses) has thus far not been attended by very satisfactory results. Fairly sound conclusions may often be reached by a study of field relations, mineral composition, texture, or local effects of weathering, but it is only rarely that the chemical composition of the schist points clearly to a sedimentary or igneous origin, and even then the student feels safer when chemical evidence is confirmed by criteria of other kinds. It is purposed here to test the validity of chemical criteria in the study of the origin of schists.

Fundamental assumptions in the use of chemical criteria for the identification of schists are (1) that these rocks before rock flowage had a distinctive original composition sufficient to identify them as igneous rocks or sediments, and (2) that schists have retained essentially this composition during rock flowage, or at least that changes in composition during these processes have not been sufficient entirely to mask the original chemical composition. Both of these assumptions require consideration.

(1) In proportion as decomposition and sorting are effective in making sediments, the sediments take on chemical characteristics which serve to distinguish them from igneous rocks. Some of these features are dominance of magnesia over lime, potash over soda, excess of alumina, high silica, and in carbonate formations excessively high lime and magnesia.[1] But where sediments have been produced by disintegration of igneous rocks or without marked assortment, their composition may differ but little from that of the parent rock. Igneous rocks may be weathered or hydrothermally altered to such an extent that they approach the composition of sediments. It may be noted that the shales, which are quantitatively the most important sediments, are more largely changed to schists and slates than are other sediments. Shales in general show less variation from the composition of igneous rocks than do other sediments like quartz-sand or limestone, which less frequently become schistose or gneissic.

It is apparent, therefore, that while broadly there are distinctive differences in chemical composition between igneous and sedimen-

[1] Bastin, Edson S., Chemical composition as a criterion in identifying metamorphosed sediments: Jour. Geol., vol. 17, 1909, p. 449.

tary rocks, there are large quantities of igneous rocks which may not be identified as such by chemical determination alone, that is, igneous rocks which are weathered or hydrothermally altered, and that there are extensive deposits of sediments, such as those formed by disintegration in arid climates and in many unassorted continental deposits, which fail to show distinctive sedimentary composition. Therefore even if original composition were retained during the process of schist making, and the schist were found to have distinctively igneous or distinctively sedimentary composition, it is impossible to be certain whether the original rock was igneous or sedimentary.

(2) The second common assumption in the use of chemical criteria for the identification of schists and gneisses is that during the process of schist-making, the composition has not essentially changed, or at least that the changes in composition during these processes have not been sufficient to entirely mask the original chemical composition.

This assumption is based essentially on the belief that the conditions of anamorphism or rock flowage are not likely to favor transfer of materials, rather than on quantitative proof that the composition has not been changed. There are surprisingly few cases of igneous or sedimentary rocks which have been traced, step by step, with quantitative analyses, into schists and gneisses, and the few cases available do not adequately support the assumption of maintenance of composition during anamorphism or rock flowage. A shale or mud passes into a slate or schist with loss of water, carbon dioxide, and oxygen. The loss of such substances as water, carbon dioxide, oxygen, and even lime, during anamorphism is ordinarily accepted as a matter of course. The proposition that in some cases there may be elimination of silica, or any other substance present in excess of the requirements of the schist-making minerals, is not so well established. It is a fact, none the less, that analyses often show loss of silica where a quartzite changes to a sericite schist, and that a limestone or dolomite may change to a talc schist or other silicate rock with considerable loss of carbon dioxide and lime. The well known growth of platy and columnar minerals like mica, chlorite, and hornblende in schists, with corresponding diminution in proportion of other minerals, especially feldspars, affords a contrast mineralogically between schists and

parent rocks, which favors the view that there has been some change in chemical composition. The new minerals characteristic of schists do not in many cases seem to require all of the substances of the original rock. On other pages we have attempted to show a convergence in schist composition toward that of the platy or columnar minerals principally causing the schistosity.

Quantitative evidence is certainly sufficient to establish some degree of change and to a certain extent the nature of the change, in processes of rock flowage or anamorphism. It is not sufficient to show clearly the extent of this change, but certainly it raises serious question as to the correctness of the assumption that there is no essential change in composition during rock flowage or anamorphism and should make us cautious in assuming the composition of a schist or gneiss to be essentially that of the parent rock.

Triangular diagrams. To test chemical criteria in specific cases, analyses of common schists, both of known and unknown origin, are entered on the accompanying diagrams, with a view to bringing out the essential chemical features of these schists and the possibility of using them in determining origin. Analyses are platted in such a fashion as to bring out the distinctive differences between sedimentary and igneous characters. The distinctive features of sedimentary origin are dominance of magnesia over lime, potash over soda, excess of alumina, and high silica.[1] The distinctive features of igneous origin are conversely lower alumina, dominance of soda over potash, and lime over magnesia. On each diagram the numbered points represent separate analyses, the numbers referring to the schedule of analyses which accompanies each diagram. The position of the point in the triangle shows the percentage of silica, measured toward the apex, the percentage of alumina, measured toward the lower left hand corner, and the sum of the percentages of other comstituents, measured toward the lower right hand corner. In order to economize space and secure a larger scale, the triangles used are, in reality, portions of much larger triangles. In each of them the percentage of silica for any point may be read from the numbers along the right hand slope of the triangle and the

[1] Bastin, Edson S., Chemical composition as a criterion in identifying metamorphosed sediments: Jour. Geol., vol. 17, 1909, p. 449.

percentage of alumina along the left hand slope. In order to compare the silica-alumina ratio in each of the analyses with the silica-alumina ratio in the related rocks and minerals, broken lines are drawn across the triangle, indicating the ratio of silica to alumina in the most important rocks and minerals genetically related to the particular group of analyses considered. Other characteristics of each analysis are shown as follows: The difference in percentage of lime and magnesia is measured by a horizontal line drawn from the point representing the analysis, the excess of lime over magnesia being shown by a line to the right, and the excess of magnesia over lime by a line to the left. The difference in percentage between soda and potash is represented by an inclined line running downward to the right or upward to the left, the excess of potash over soda being shown by a line extending downward to the right, and an excess of soda over potash by a line extending upward to the left. The excess or deficiency of alumina as compared with the amount required to satisfy the ratio of alumina to soda, potash, and lime, in the feldspars, is shown by a line extending upward to the right or downward to the left. If the percentage of alumina in the rock exceeds by 5 per cent the amount required to give a molecular ratio equaling the sums of the molecular ratios of soda, potash, and lime, the alumina is represented as showing neither excess nor deficiency. If the alumina present in the analysis exceeds by more than 5 per cent the amount of alumina required to yield a molecular ratio, equal to the sum of the molecular ratios of soda, potash, and lime, then the alumina is considered to be in excess and this excess is indicated by a line drawn downward to the left, each unit of the large triangular scale being taken as two units of alumina. If the percentage of alumina in the rock is less than 5 per cent in excess of the amount of alumina required to yield a molecular ratio, equal to the sums of molecular ratios of soda, potash, and lime, the line is drawn upward to the right, each unit division of the large triangle representing two units of alumina.

It will be seen that lines representing excess of soda over potash, the deficiency of alumina, and excess of lime over magnesia, all three of which are characteristic of igneous rocks, are grouped on one side of the point, while the characteristics of sedimentary rocks, namely, excess of magnesia over lime, excess of alumina,

and excess of potash over soda, are grouped on the opposite side of the point representing an analysis. By connecting the ends of the lines of each group, as shown in the key diagram accompanying each of the triangular diagrams, small triangular pennants are formed, which make it possible easily to recognize either the igneous or sedimentary characteristics of analyses. For ex-

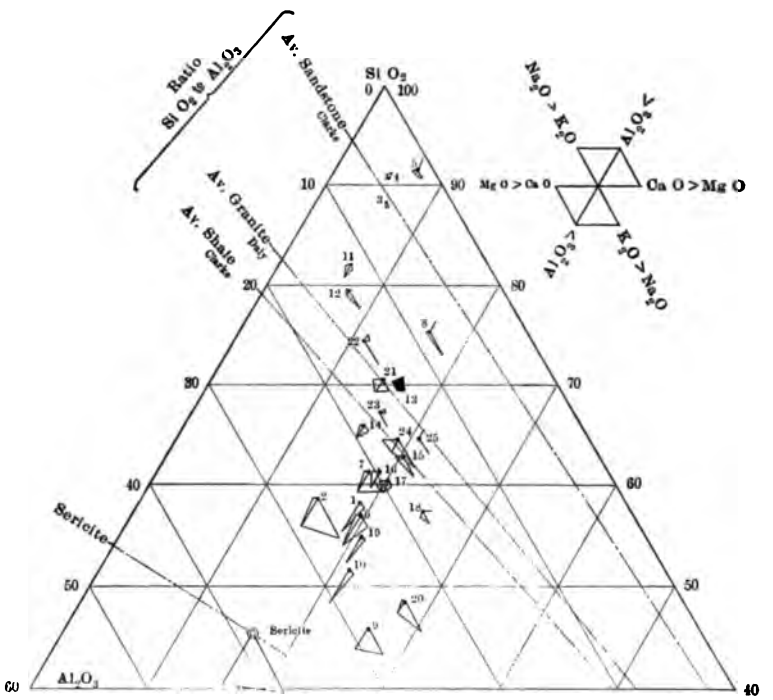


FIG. 14. Analyses of sericite schists platted to show their igneous and sedimentary characteristics. Compare with Figs. 15 and 16.

ample, in the triangular diagram of the sericite schists, analysis No. 2 has all three of the common ratios of sediments, while analysis No. 13 has all three of the chemical ratios ascribed to igneous rocks.

Fig. 14. Sericite schists. On this diagram 25 sericite schist analyses are represented. The list follows:

1. Developed from Waterloo quartzite near Waterloo, Wisconsin. Warner, J. H., unpublished bachelor thesis, Univ. of Wis.

2. 3. Developed from Mesnard quartzite, Marquette district, Michigan.
4. Quartz-sericite-schist. Sedimentary. Daly, R. A., Bull. 209, U. S. Geol. Survey, p. 15.
5. Quartz-sericite-schist. Probably sedimentary. Clarke, F. W., Bull. 228, U. S. Geol. Survey, p. 59, D.
6. Phyllite. Sedimentary. Clarke, F. W., Bull. 228, U. S. Geol. Survey, p. 60, F.
7. Sericite schist developed from slate near granite contact. Hawes, G. W., Bull. 330, U. S. Geol. Survey, p. 533.
8. Mica schist. Probably sedimentary. Teall, J. J. H., The geological structure of the northwest Highlands of Scotland, Memoir Geol. Survey Gt. Britain, 1907, p. 76.
9. Mica schist. Probably sedimentary. The geological structure of the northwest Highlands of Scotland, Memoir Geol. Survey Gt. Britain, 1907, p. 80.
10. Sericite schist from Georgia. Unknown origin. Steidtmann, Edward, unpublished analyses.
11–20, inclusive. Sericite schists. Origins not given. Kemp, J. F., Handbook of Rocks, p. 127.
21. Sericite schist developed from an acid tuff, Marquette district, Michigan. Bayley, W. S., Mon. 28, U. S. Geol. Survey, p. 168.
22. Sheared granite from Pikes Peak district. Mathews, E. B., Bull. 228, U. S. Geol. Survey, p. 161.
23. Sheared quartz-porphyry. Williams, G. H., Bull. 62, U. S. Geol. Survey, p. 121.
24–25. Sericite schists. Origin not given. Pirsson, L. V., Rocks and rock minerals, p. 363.

These analyses are selected as typical of sericite schists from many localities. Of the analyses platted 1, 2, 3, 4, 6, 7, and 21, have been described as being of sedimentary origin; 5, 8, 9, as probably sedimentary, 10 and 11 to 20 inclusive, and 24 and 25 either unknown or origin not given, and 22 and 23 have been described as of igneous origin. Of the 25 analyses only one, No. 13, has all of the chemical characteristics of an igneous rock. Nine of the analyses have all the three principal qualities ascribed to sedimentary rocks and the remaining fifteen are divided between igneous and sedimentary characteristics.

One of the most striking features of the diagram is the fact that over half of the analyses platted show a silica-alumina ratio lower than that of the average rocks from which sericite schists are characteristically derived, namely, sandstone, granite, and shale. There seems to be a distinct tendency for the sericite

schist to approach the composition of the mineral sericite, which is shown by a point on the diagram.

Figs. 15 and 16. Acid igneous rocks altered by weathering and hot solutions. Introduced for comparison with sericite schists. It is of interest to compare the diagram for sericite schists with that of acid igneous rocks altered by surface weathering and acid igneous rocks altered by hot solutions.

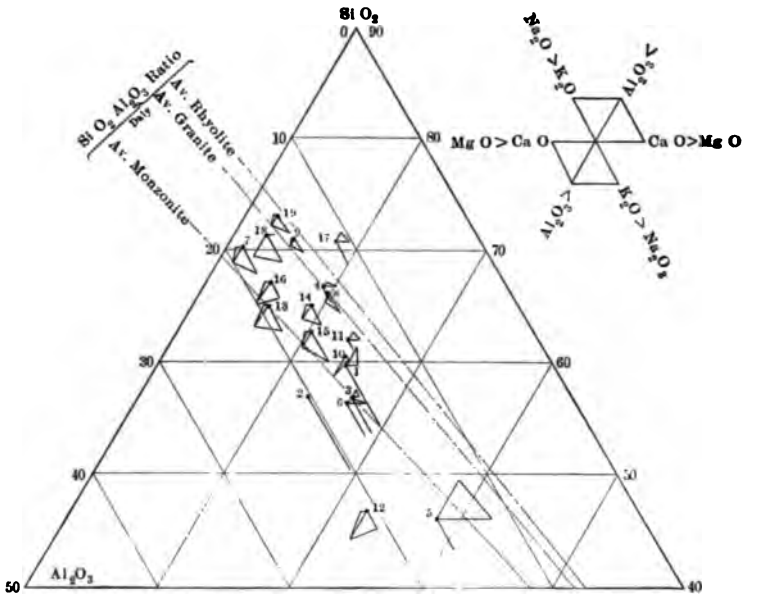


FIG. 15. Analyses of hydrothermally altered acid igneous rocks platted to show igneous and sedimentary characteristics. Compare with Fig. 14.

In Fig. 15, 19 analyses of hydrothermally altered acid igneous rocks are shown.

1. Granitic rock, Cripple Creek, Colorado. Lindgren, W., and Ransome, F. L., Prof. Paper 54, U. S. Geol. Survey, p. 194.
2. Latite phonolite, Cripple Creek, Colorado. Lindgren, W., and Ransome, F. L., Prof. Paper 54, U. S. Geol. Survey, p. 189.
3. Monzonite porphyry, Bingham district, Utah. Boutwell, J. M., Prof. Paper 38, U. S. Geol. Survey, p. 178.
4. Granitic rock, Willow Creek district, Idaho. 18th Ann. Rept., U. S. Geol. Survey, pt. 3, p. 640.

5. Granodiorite, Placer County, California. Trans. Am. Inst. Min. Engrs., vol. 30, 1901, p. 666.
6. Granodiorite, Nevada County, California. Trans. Am. Inst. Min. Engrs., vol. 30, 1901, p. 666.
7. Granite, Pilbara gold field, Australia. Lindgren, W., Econ. Geol., vol. 1, 1906, p. 540.
8. Quartz monzonite, Rimini, Montana. Knopf, Adolph, Econ. Geol., vol. 8, 1913, p. 115.
9. Rhyolite porphyry, Butte, Montana. Weed, W. H., Prof. Paper 74, U. S. Geol. Survey, p. 90.

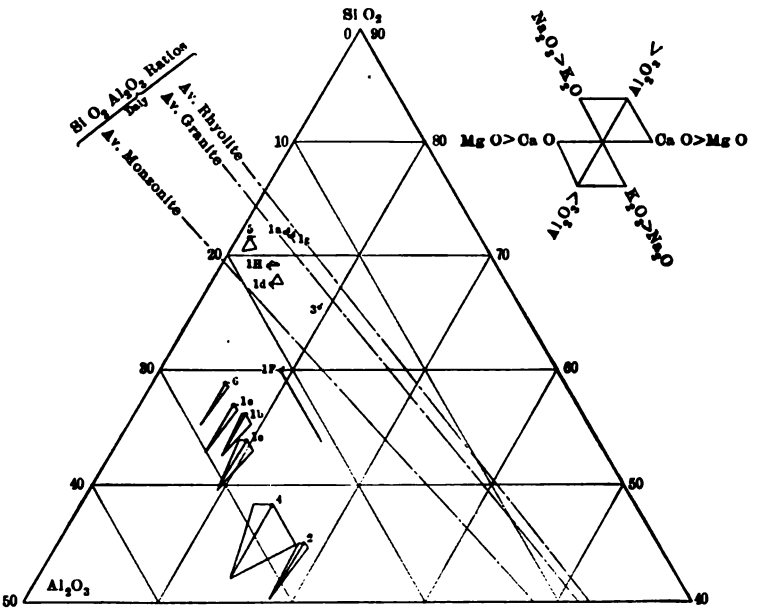


FIG. 16. Analyses of weathered acid igneous rocks platted to show igneous and sedimentary characteristics. Compare with Figs. 14 and 15.

10. Dacite, Goldfield, Nevada. Ransome, F. L., Prof. Paper 66, U. S. Geol. Survey, p. 181.
11. Hornblende dacite, Hauraki gold field, New Zealand, Finlayson, A. M., Econ. Geol. vol. 4, 1909, p. 638.
12. Monzonite porphyry, Clifton-Morenci district, Arizona. Lindgren, W., Prof. Paper 43, U. S. Geol. Survey, p. 168.
13. Quartz monzonite, Butte, Montana. Weed, W. H., Prof. Paper 74, U. S. Geol. Survey, p. 88.
14. Granite, Pittsmont Mine, Butte, Montana. Unpublished.
15. Quartz monzonite, San Francisco district, Utah. Butler, B. S., Prof. Paper 80, U. S. Geol. Survey, p. 76.

16. Quartz monzonite, San Francisco district, Utah. Butler, B. S., Prof. Paper 80, U. S. Geol. Survey, p. 76.
17. Porphyry, Rio Tinto, Spain. Lindgren, W., Mineral Deposits, p. 604.
18. Monzonite, Tintic, Utah. Lindgren, W., Mineral Deposits, p. 525.
19. Hornblende-mica andesite, Tonopah, Nevada. Lindgren, W., Mineral Deposits, p. 452.

Of these nine have the three principal chemical features of sedimentary rocks, while none of the analyses have all three of the principal features of igneous rocks.

In Fig. 16 thirteen analyses of weathered acid igneous rocks are shown.

1a–1h. Granite gneiss, Georgia. Watson, T. L., Granites and gneisses of Georgia.
2. Gneiss micaceous, Virginia. Merrill, G. P., Bull. Geol. Soc. Am., vol. 8, 1897, p. 160.
3. Granite, District of Columbia. Merrill, G. P., Bull. Geol. Soc. of Am., vol. 6, 1895, p. 323.
4. Granite, Scotland. Holland, P., and Dickson, E., Proc. Liverpool Geol. Soc., vol. 7, pt. 1, p. 116.
5. Granite, Montana, one-quarter of a mile southeast of Homestake. Kirk, C. T., unpublished.
6. Syenite, Arkansas. Williams, J. F., Ann. Rept. Geol. Survey Arkansas, 1890, vol. 2, pp. 82 and 88.

Of these analyses four have all of the three principal characteristics attributed to sedimentary rocks, while none of them have the three principal characteristics attributed to igneous rocks.

It obviously follows from this that a sericite schist developed from a hydrothermally altered acid igneous rock, or, from the weathered portion of an acid igneous rock, might easily be mistaken for a sedimentary schist from chemical evidence alone.

Chlorite schists, Fig. 17. On Fig. 17 are 11 analyses of chlorite schist. The list follows:—

1. Chlorite schist. Origin not given. Kemp, J. F., Handbook of Rocks, p. 131.
2. Chlorite schist. Igneous. Bascom, F., Bull. 136, U. S. Geol. Survey, p. 78.
3. Chlorite schist. Origin not given. Pirsson, L., Rocks and Rock Minerals, p. 376.
4–7, inclusive. Chlorite schists developed from greenstones, etc. Williams, G. H., Bull. 62, U. S. Geol. Survey.

8. Chlorite schist. Sedimentary origin. Country rock at Morro Velho Mine, Minas Geraes, Brazil. Shearer, H. K., unpublished analyses.
9. Schistose lava from Port Isaac. Teall.
10. Chloritic gneiss. Igneous. Clarke, F. W., Bull. 228, U. S. Geol. Survey, p. 57.
11. Epidote chlorite schist. Origin not given Clarke, F. W., Bull. 228, U. S. Geol. Survey, p. 63.

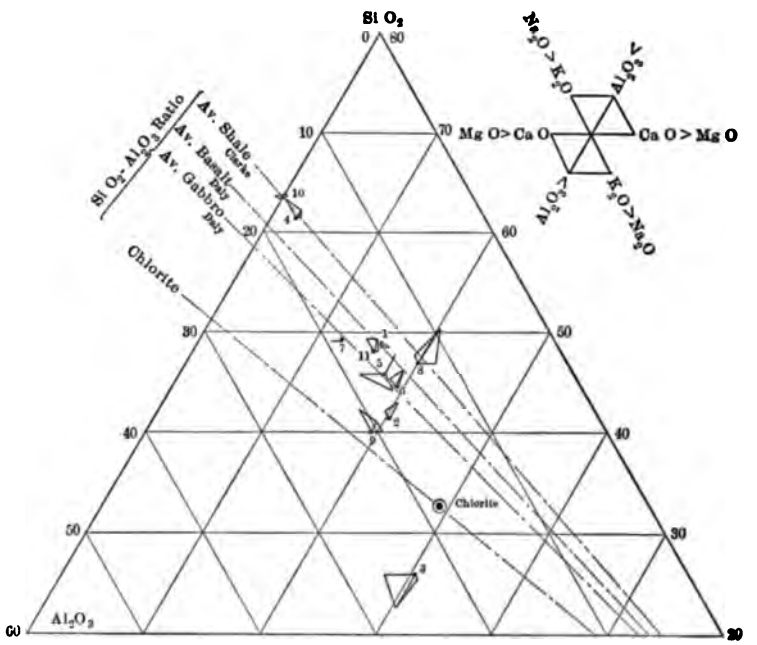


Fig. 17. Analyses of chlorite schists platted to show their igneous and sedimentary characteristics. Compare with Figs. 19 and 20.

Of these, two have the three principal characteristics of sedimentary rocks; five have the three principal features of igneous rocks, and the remainder are divided. Note that analysis No. 10, which has the composition of a sedimentary rock, represents a schist which was developed from a granite and that analysis No. 8, which shows the three principal characteristics ascribed to igneous rocks, is a chlorite schist developed from a sedimentary shale. As in the case of sericite schists there is convergence in

composition toward the dominant or characteristic mineral, in this case chlorite.

Hornblende schists, Fig. 18. On Fig. 18 are 15 analyses of hornblende schists and amphibolites from sources given in the following list:

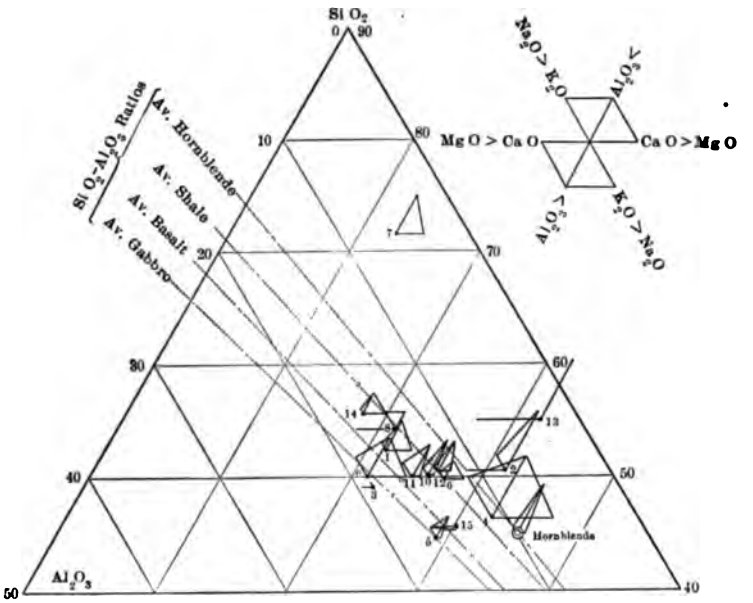


FIG. 18. Analyses of hornblende schist and amphibolites platted to show their igneous and sedimentary characteristics. Compare with Figs. 19 and 20.

1–5, inclusive. Kemp, J. F., Handbook of Rocks, p. 130. No. 2 is regarded by Becker as a metamorphosed sediment. Kemp believes others to be igneous.

6. Hornblende schist developed from diabase dike. The geological structure of the northwest Highlands of Scotland: Memoir Geol. Survey Gt. Britain, 1907, p. 95.
7. Quartz-hornblende schist. Probably sedimentary. The geological structure of the northwest Highlands of Scotland: Memoir Geol. Survey Gt. Britain, 1907, p. 80.
8. Amphibolitic schist, metamorphosed tuff. Clarke, F. W., Bull. 228, U. S. Geol. Survey, p. 251, E.
9. Amphibolite, described as metamorphosed limestone. Adams, F. D.,

and Barlow, A. E., Geology of the Haliburton and Bancroft areas, Ontario: Mem. 6, Geol. Survey Canada, 1910, p. 104.

10. Hornblende schist, Vestana, Sweden. Unknown origin. Pirsson, L. V., Rocks and Rock Minerals, p. 344.
11. Schistose amphibolite, Whitman's Ferry, Sunderland, Massachusetts. Unknown origin. Pirsson, L. V., Rocks and Rock Minerals, p. 381.
12. Amphibolite, Crystal Falls district, Michigan. Igneous. Mon. 36, U. S. Geol. Survey, p. 397.
13. Amphibolite, Chiavenna. Unknown origin. Pirsson, L. V., Rocks and Rock Minerals, p. 381.
14. Amphibolite, Goshen, Massachusetts. Probably derived from limestone. Pirsson, L. V., Rocks and Rock Minerals, p. 381.
15. Amphibolite, pyritiferous, Ophir district, California. Described as sedimentary in origin. Pirsson, L. V., Rocks and Rock Minerals, p. 381.

Of these analyses numbers 2, 7, 8, 9, 14, and 15 have been described as being definitely or probably of sedimentary origin. Numbers 6 and 12 are of igneous origin. The remainder are of unknown origin. Analyses 9 and 14, which are described as metamorphosed limestone, and No. 15, which is also of sedimentary origin, show the three special features of igneous rocks. None of the analyses have all three common features of sedimentary rocks, although several show excess of MgO over CaO.

The diagram brings out the fact that the hornblende schists tend to approach hornblende in composition. This is shown both by the position of the points and the shape of the pennants. The fact that none of the analyses of hornblende schists have all of the chemical ratios of sediments is related to the fact that hornblende itself has all three of the characteristics of igneous rocks.

Figs. 19 and 20. Basic igneous rocks altered by weathering and by hot solutions. Introduced for comparison with chlorite schists and hornblende schists.

In Fig. 19, 11 analyses of hydrothermally altered basic igneous rocks are shown.

1. Diorite, Willow Creek district, Idaho. 20th Ann. Rept. U. S. Geol. Survey, pt. 3, p. 219.
2. Diorite prophyry, Breckenridge district, Colorado. Ransome, F. L., Prof. Paper 75, U. S. Geol. Survey, p. 96.

3. Amphibole schist, Placer county, California. Trans. Am. Inst. Min. Engrs., vol. 30, 1901, p. 666.
4. Diabase, Grass Valley, California. Trans. Am. Inst. Min. Engrs., vol. 30, 1901, p. 666.
5. Amphibole schist, Kalgoorlie, W. Australia. Lindgren, W., Econ. Geol., vol. 1, 1906, pp. 536–537.
6. Hornblende andesite, Hauraki gold field, New Zealand. Finlayson, A., Econ. Geol., vol. 4, 1909, p. 637.

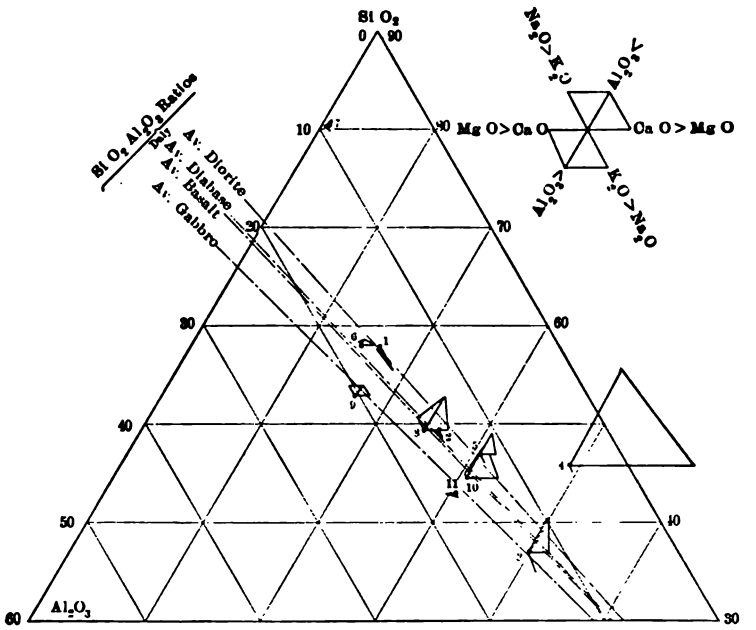


FIG. 19. Analyses of hydrothermally altered basic igneous rocks platted to show igneous or sedimentary characteristics. Compare with Figs. 17, 18, and 20.

7. Diabase, Comstock Lode. Becker, G. F., Mon. 3, U. S. Geol. Survey, p. 153.
8. Trap (melaphyre), Keweenaw Point. Van Hise, C. R., and Leith, C. K., Mon. 52, U. S. Geol. Survey, 1911, p. 583.
9. Amphibolite, Alaska. Knopf, Adolph, Bull. 502, U. S. Geol. Survey, p. 39.
10. Diorite, Gold Creek, Alaska. Lindgren, W., Mineral Deposits, p. 646.
11. Andesite, Tonopah, Nevada. Lindgren, W., Mineral Deposits, p. 454.

Of these, four have the three principal characteristics of igneous rocks; the remaining seven are divided, none of them showing the common ratios of sedimentary rocks.

On Fig. 20 are platted 13 analyses of weathered basic igneous rocks.

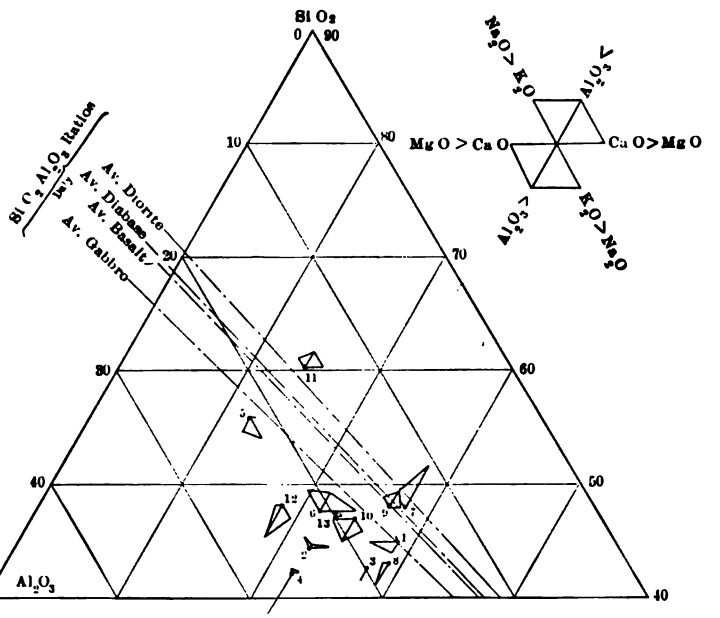


Fig. 20. Analyses of weathered basic igneous rocks platted to show igneous or sedimentary characteristics. Compare with Figs. 17 and 18.

1. Diabase, Island of Jersey. Holland, P., and Dickson, E., Proc. Liverpool Geol. Soc., vol. 7, 1892–93, p. 108.
2. Diabase, Medford, Massachusetts. Merrill, G. P., Bull. Geol. Soc. of Am., vol. 7, 1896, p. 353.
3. Diabase, Spanish Guiana, Venezuela. Clarke, F. W., Bull. 330, U. S. Geol. Survey, p. 413.
4. Diorite, Virginia. Merrill, G. P., Rocks, Rock Weathering and Soils, 1906, p. 225.
5. Phonolite, Bohemia. Merrill, G. P., Rocks, Rock Weathering and Soils, 1906, p. 198.
6. Gabbro, Allen Junction, Minnesota. Unpublished analyses.
7. Diabase, Chatham, Virginia. Watson, T. L., Am. Geologist, vol. 22, 1898, p. 87.
8. Basalt, Bohemia. Merrill, G. P., Rocks, Rock Weathering and Soils, 1906, p. 223.

9. Trap rock, Connecticut Valley. Hawes, G. W., Am. Jour. Sci., vol. 9, 1875, p. 185.
10. Dolerite, Rowley Regis, England.
11. Augite andesite, Washington.
12. Diabase, Penokee-Gogebic district, Michigan.
13. Augite diorite, Magnetberg, Southern Urals. Clarke, F. W., Bull. 330, U. S. Geol. Survey, p. 413.

Of these two have the three principal sedimentary ratios; three have the three principal igneous ratios, and the remaining eight are divided.

Chlorite or hornblende schists, derived from these rocks might well show either sedimentary or igneous composition.

Conclusion as to use of chemical criteria. Chemical criteria do not satisfactorily discriminate schists of sedimentary and igneous origin. Schists of known sedimentary origin are in some cases shown to have the composition ordinarily ascribed to igneous rock, and schists of known igneous origin in some cases would be classed as sedimentary by the use of these criteria. Chemical criteria of origin have value only when carefully qualified and limited. As it is often difficult to discriminate the qualifying factors, the use of chemical criteria must be a very tentative one. They do not apply where there has been weathering of igneous rocks or where there has been thermal alteration of igneous rocks, or where there has been contact alteration of sediments, or where, during dynamic metamorphism, there has been considerable change in composition, as seems to be the case in certain quartzites and limestones, which originally have varied widely from the composition of the schists. In rocks which are so far metamorphosed as to require chemical criteria for identification of origin, it is practically an impossibility to eliminate in many cases the possibilities of earlier alteration above mentioned.

In other words, in the very cases where other criteria fail, the chemical data likewise fail. Conversely, chemical criteria may warrant decisive classification only in cases where other criteria are decisive and chemical criteria are thus not needed.

General Discussion of Methods of Identifying Gneisses and Schists

The writers know of no case where all the evidences above cited have been used in the determination of sedimentary origin

of a gneiss. As one surveys the methods and conclusions in various investigations of gneisses and schists, it becomes apparent that no one criterion is ordinarily sufficient to establish sedimentary origin. Chemical composition has been used with some confidence, yet sedimentary origin has been argued from field relations and mineralogical composition for some gneisses which have chemical composition of igneous rocks. Two other methods are likely to receive more attention in the future, namely, the study of heavy residuals and determination of temperature of crystallization of the quartz.

Gneisses developed secondarily from igneous rocks by pressure and recrystallization have been positively identified in even fewer cases than sedimentary gneisses. Many gneisses are known to be original igneous rocks with flow structure; a few have been found to be the result of mechanical breaking down by granulation, for instance, the granulated anorthosites near Montreal.[1] Many gneisses have been described as the foliated equivalents of granites as result of pressure and recrystallization, but often without adequate proof of this relation. Lehmann [2] has apparently shown the development of gneisses from granites in the Saxony area. In parts of the Lake Superior country there are gneisses which seem to have such relations to granite gneisses as would result from secondary pressures and recrystallization, but there is not a single proved case there. Many pairs of analyses of granites and equivalent gneisses have been published, but these have usually been made on the assumption that the gneiss was the result of secondary alteration of granite and without adequate consideration of the possibility that gneissose structure may be an original flow structure.

Many more schists than gneisses have been proved to be the result of mashing of igneous rocks, for instance, the chlorite schists so commonly developed from the mashing of basalt, the hornblende schists formed in these rocks by contact metamorphism of granites, and micaceous schists formed in granites and porphyries along a shear zone.

[1] Adams, F. D., Report on the geology of a portion of the Laurentian area lying to the north of the island of Montreal: Ann. Rept., Geol. Survey Can., vol. 8, pt. J, 1896, pp. 85 et seq.

[2] Lehmann, J., Enstehung der Altkrystallinischen Schiefergesteine, Bonn, 1884. Referred to by Harker, A., Petrology for students, 1895, p. 285.

This brings us back to the suggestion made on an earlier page, that when igneous rocks break down by mashing in the presence of water, there tend to develop the platy and columnar hydrous minerals characteristic of schists. These minerals are the same in kind as those derived from the anamorphism of a sediment. As compared with the igneous rock, the change to a schist amounts to katamorphism, and requires the introduction of water and carbon dioxide. To whatever extent gneiss may be formed by the mashing of igneous rock, and, as noted, this extent is problematical, conditions different from those forming schists are implied by the fact that the gneisses have relatively less amounts of platy and columnar hydrous minerals. We have suggested that gneisses may form only in places where the agents of hydration and carbonation are lacking, and that where these agents were present, the change is more toward the schist type.

The terms "schist" and "gneiss" have been used as representing two types of rocks. It is of course to be recognized that there are complete gradations between schist and gneiss; that it probably follows, therefore, that there are many conditions of origin of the schists and gneisses from igneous rocks intermediate between those described.

CHAPTER II

OCEAN, LAKE, RIVER, AND UNDERGROUND SOLUTIONS AS BY-PRODUCTS OF THE METAMORPHIC CYCLE

Water is the solvent and carrier in most of the chemical changes of metamorphism. The mineral content of surface and underground waters should therefore bear some relation to the chemical changes in the rocks with which they have come into contact. As a matter of fact, they afford an interesting check on conclusions as to the nature and extent of the metamorphic reactions inferred from study of altered rocks.

DISSOLVED MINERAL CONTENT OF UNDERGROUND WATERS

Source and its Effect on Proportions of Dissolved Mineral Substances

It is proposed here to discuss the dissolved mineral content of underground waters in its relation to the metamorphic processes in the associated rocks. Some reference has been made to the distribution and flow of underground waters, but since the emphasis is on the inductive study of rock alterations no general treatment of this subject is attempted. For a general discussion of the flow and work of underground waters the reader is referred to Van Hise's Monograph on Metamorphism.

Some underground waters of supposedly magmatic or juvenile origin are believed to carry, directly from a magma, substances quite different from those furnished by processes of rock alteration. In the discussion of alterations by hot waters in preceding chapters mention has been made of the local contribution, from magmatic sources, of certain metallic ores, sulphur, boron, fluorine, and other constituents.

The principal dissolved constituents of underground waters are derived from the secondary alterations of rocks with which they

come in contact. The contributions from these sources vary with the kinds of rock, with the conditions and degree of their metamorphism, with the temperature of the water, its content of acids or salts, and with other factors. The most abundant dissolved constituents in the waters within our range of observation result from rock weathering. Cementation and dynamic metamorphism may add to or subtract from the dissolved mineral content of underground waters. This effect, however, is a minor one in most cases, not easily identified or measured, but rather inferred from a study of the rock alterations. The relation of the dissolved mineral content of natural waters to metamorphic processes may best be studied by first comparing the dissolved mineral content of the waters with the contributions by weathering, and then treating as modifying factors the effects of precipitation, cementation, dynamic metamorphism and other metamorphic processes.

Weathering of acid or basic igneous rocks under given conditions yields solutes in fairly well determined proportions. Analyses of underground waters check out fairly well with the contributions inferred by comparing analyses of fresh and weathered igneous rocks, when allowance is made for modifying factors discussed below. The principal reason for the differences between the mineral content of underground waters and the contributions inferred from comparison of fresh and altered rocks is to be found in the differing rates at which substances are contributed to solution by weathering. If the several substances lost by weathering were contributed to the solutions with the same speed, the proportions in the ground waters should check closely with those inferred from a comparison of analyses of fresh and weathered rocks. Certain substances go into solution rapidly and others very slowly, hence the composition of waters at any one time may differ widely from the composition inferred from total losses during weathering. Early in the process of rock decay the underground waters may show a higher percentage of the more soluble materials, while later the less soluble materials are relatively increased. At no one time do the proportions in solution correspond exactly to the proportions of the several constituents in the total losses from the weathered rock of a given area.

If the waters are hot, as in certain volcanic areas, the proportions of substances contributed to the solutions have certain

characteristics differing from those of normal weathering. (See pp. 45 et seq.)

Underground waters in contact with sediments are likely to show relatively high lime and magnesia, as these substances occur in the sediments as easily soluble carbonates. So far as the fragmental sediments contain undecomposed igneous rock materials, they yield proportions of substances not far different from the contributions of weathering igneous rocks. So far as they consist of the products of advanced decomposition they yield difficultly soluble substances, as would the residual products of rock decay in place.

Saline deposits yield salt water and brines. Rocks containing sulphides yield sulphates of the bases.

Many interesting cases of the relations of mine waters to associated wall rock and ore alterations might be cited. Conspicuous illustrations are the content of sulphate of copper derived from oxidation of copper sulphides, and the high silica content of waters percolating through the iron formations of the Lake Superior iron districts. The mineral content of mine waters is of course widely changed by secondary precipitation before the water joins the runoff, although in connection with sulphide deposits enough of the sulphates remain in solution and join the runoff to materially affect the composition of the waters draining a sulphide region.

Additional Factors Modifying the Proportions of Dissolved Mineral Substances

The process of cementation modifies the mineral content of underground waters, principally by the abstraction of silica, lime and magnesia carbonates, and iron salts. This tends to increase the proportion of the other constituents present, principally soda and potash. Potash is believed to be removed from the water to some extent through adsorption by clay. Magnesia is abstracted from solution more than lime with the result that the lime content of underground waters is likely to be higher than that indicated by weathering.

Dynamic metamorphism modifies the content of underground waters by contributing to it lime, carbon dioxide, oxygen, and in some cases silica. These contributions are probably small as compared with the contributions from weathering.

Underground waters carry more iron than do surface waters, because they afford less opportunity for oxidation. This is evidenced by iron deposits about springs, where the water has been precipitated by oxidation on coming to the surface.

The amount and speed of flow of the water affects the mineral content. The total amount of solution accomplished by rapidly moving waters of low salinity may be much greater than by slowly moving waters of high salinity, hence, the degree of salinity is not in proportion to the work done. Where there is abundant water moving rapidly the salinity is less than where the water is less or moving more slowly. This affects also the relative proportions of substances in solution. Deep waters moving slowly are likely to have a relatively higher content of soda and lime chlorides. This is true whether the rocks are igneous or sedimentary and more or less regardless of the nature of the sedimentary rock. For instance, such waters are found in the deep copper mines of Michigan, in the deep iron mines of Michigan, the Paleozoic sediments of Wisconsin and northern Illinois, and in the granites of the Piedmont area of Georgia. They have been called fossil sea waters or connate waters when found in sedimentary rocks, and taken to represent the composition of the early sea.[1] In some cases these have been ascribed to magmatic sources. Salt waters in shallow porous strata near the ocean have been traced to direct percolation from the sea. The very general and widespread similarity of the salt waters in deep mines and borings, both in igneous and sedimentary rocks, strongly suggests that they owe their character principally to being confined without flowage or renewal, with the result that the more soluble substances, like soda, lime, and chlorine, greatly increase while the carbonates are precipitated. In other words, all the chlorine which gets into solution from any source stays in, while larger amounts of carbonates may have come into solution but have been precipitated. The controlling principles seem to be not far different from those which determine the high chlorine content of oceans and enclosed surface basins described later in this chapter. Even river waters at the surface show an increase of chlorine toward their mouths as compared with carbon dioxide, due to the fact that all of the chlorine which gets into the water

[1] Lane, A. C., Mine waters: Proc. Lake Sup. Min. Inst., vol. 13, 1908, pp. 63 et seq.

stays there, while the carbon dioxide (as carbonates) may be precipitated and redissolved many times, and may be replaced by chlorine or the sulphate radical.

Conclusion as to Mineral Content of Underground Waters

In general, then, the content of underground waters is determined principally by the substances contributed by weathering, but the amounts and proportions are modified by the differing speed of solution of different minerals, effects of cementation, dynamic metamorphism, degree of advancement of weathering, amount and flow of underground waters, temperature of the waters, and other factors. The differing rates of solution of substances under weathering determine to a large extent their proportion in the dissolved mineral matter in a given water. Cementation tends to abstract certain substances more than others. A fact which stands out in study of underground waters is that the mineral content is determined to a larger extent by the solubility of carbonates in sediments than by the weathering of igneous rocks.

The mineral content of underground waters, as found in springs, wells, borings, and mine openings, probably shows a closer accord with the metamorphic changes in the rock than do the river waters. They have to some extent escaped loss of identity through dilution and mixing with water from other sources. The shallower underground water has the same relation to the rocks through which they flow as do river waters to the rocks of their drainage area.

DISSOLVED MINERAL CONTENT OF RIVERS

Source

The dissolved mineral contents of rivers are essentially soluble products of weathering. Mineral matter in magmatic solutions and constituents eliminated from rocks during anamorphism may be contributed in part to the streams, but these sources of mineral matter of rivers are unimportant as compared with weathering and cannot be satisfactorily discriminated.

Causes of Variation in Composition of River Waters

The composition of river water depends mainly on the contribution from the ground waters of its drainage basin, the losses by

precipitation during their transportation, and the amount o water, speed of flow, evaporation, and other factors effecting salin ity. These factors are of course influenced in turn by climate anc topography, as well as by all the conditions effecting the composi tion of underground water previously noted.

Not all of the materials removed in solution during the decompo sition of rocks reach the streams, and of the solutes which reacl the rivers a portion is precipitated before reaching the sea. Solu tions which reach the streams through underground courses deposi part of their dissolved material as cements, largely siliceous anc calcareous. Iron may be oxidized and precipitated soon afte solution. Silica in solution as soluble alkaline silicates may be dis placed by stronger acids and as colloidal silica be more easily precipitated. Colloidal silica is precipitated by electrolites, anc it is probable that some silica carried in river waters is precipitatec in this manner. Calcium carbonate is used to some extent by fresh water organisms, but calcareous shells are also dissolved hence the loss of lime during stream transportation is probably small and a large part of the lime and magnesia is delivered tc the sea. Potassium salts may be absorbed to some extent by the clays, probably both during the underground course of the solution and during stream transportation. Sodium suffers the least loss of any of the dissolved materials and is delivered to the sea except when deposited by desication.

The factors influencing the composition of river waters are sc interrelated that it is not easy to measure their individual effects on the mineral composition of river water. Yet, the general in- fluence of certain factors may be clearly traced. It is of interest tc compare inferred losses by katamorphism with the mineral content of rivers in specific cases where something more or less definite is known about the type of rocks of the drainage basin.

Rivers Draining Areas of Igneous Rocks

Rivers draining an area of igneous rocks show dissolved mineral content not far different from that which might be calculated from a comparison of fresh and weathered specimens of these rocks. The correspondence is not exact because of influence of other fac- tors.

The Ottawa river of Canada drains an area having a large pro-

portion of igneous rocks. By comparing analyses of fresh and altered rock, it is possible to calculate the relative amounts of materials lost and to express the materials removed in solution in terms of a total of 100 per cent. In other words, it is possible to calculate in terms of oxides the composition of the mineral matter which is removed in solution in the weathering of an igneous rock. This has been done for the acid igneous rocks and for the basic igneous rocks (platted respectively in Plates II and IV); the losses combined in a ratio of 65 parts acid and 35 parts basic are given in Table XIV.

TABLE XIV. LOSSES BY WEATHERING OF AVERAGE IGNEOUS ROCK COMPARED WITH DISSOLVED MINERAL CONTENT OF OTTAWA RIVER WATER (THE DATA ARE EXPRESSED IN TERMS OF OXIDES RECALCULATED TO TOTAL 100)

	Losses by weathering of average igneous rock	Ottawa river at Ottawa
SiO_2	63.28	30.20
Al_2O_3, Fe_2O_3	2.71	10.75
MgO	5.53	11.90
CaO	13.36	35.00
Na_2O	8.85	9.18
K_2O	6.27	2.88
Totals	100.00	99.91

In the same table the dissolved mineral matter of the Ottawa river, calculated from analysis given by Clarke,[1] is expressed in terms of oxides of the principal constituents. In Fig. 21 the mineral content of the river is compared graphically with the losses from igneous rocks. The vertical dimension of the diagram is divided to represent the ratio of mineral constituents of the Ottawa river water in terms of oxides. The shaded areas represent the ratio of constituents contributed by weathering of igneous rocks. Since the shaded areas fail to form a complete rectangle it is necessary to add to or subtract from the shaded rectangles in order to

[1] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, p. 65.

form a complete large rectangle. As the Ottawa river derives part of its solutes from other than igneous rocks, it is reasonable to assume that materials should be added to the igneous rock losses to yield the dissolved mineral matter of the river. It is evident that the material added would consist of lime, magnesia, and soda with little silica and potash.

It seems clear from comparison of Figs. 21 and 23, that the

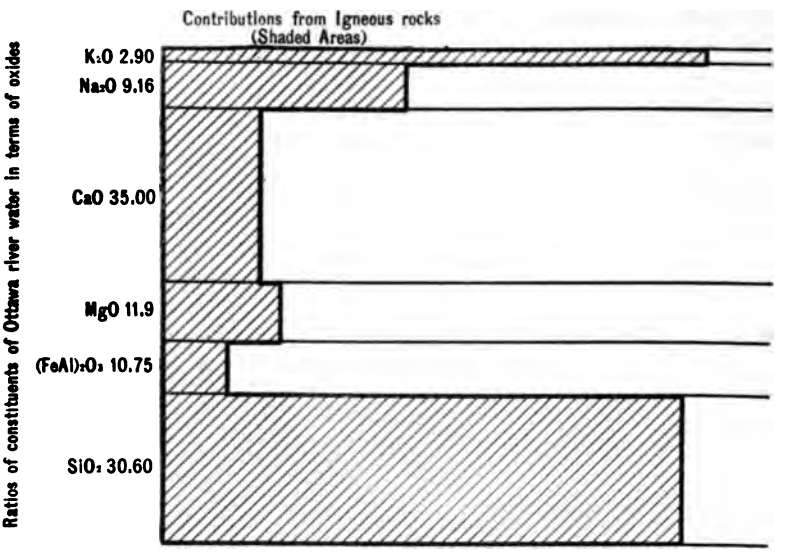


FIG. 21. Comparing graphically the losses by katamorphism of igneous rocks with the dissolved mineral content of the Ottawa river.

Ottawa river receives a larger proportion of material from igneous rocks than do the average river waters.

The Ocmulgee river above Macon, Georgia, drains an area made up practically entirely of igneous and crystalline rocks, a large part of which are granites and granite gneisses.

In Table XV the average losses by weathering of Georgia granites calculated from eight sets of analyses of fresh and weathered rocks [1] are expressed in terms of oxides. In the same table

[1] Watson, Thomas L., Granites and gneisses of Georgia: Bull. 9–A, Geol. Survey of Georgia, 1902.

TABLE XV. LOSSES BY WEATHERING OF GEORGIA GRANITE COMPARED WITH THE DISSOLVED MINERAL CONTENT OF OCMULGEE RIVER WATER (THE DATA ARE EXPRESSED IN TERMS OF OXIDES RECALCULATED TO TOTAL 100)

	Losses by weathering of Georgia granite	Ocmulgee river near Macon, Georgia
SiO_2	47.2	52.70
Al_2O_3, Fe_2O_3	3.59	2.43
MgO	3.27	4.05
CaO	15.12	17.90
Na_2O	17.00	18.28
K_2O	13.72	4.64
Totals	99.90	100.00

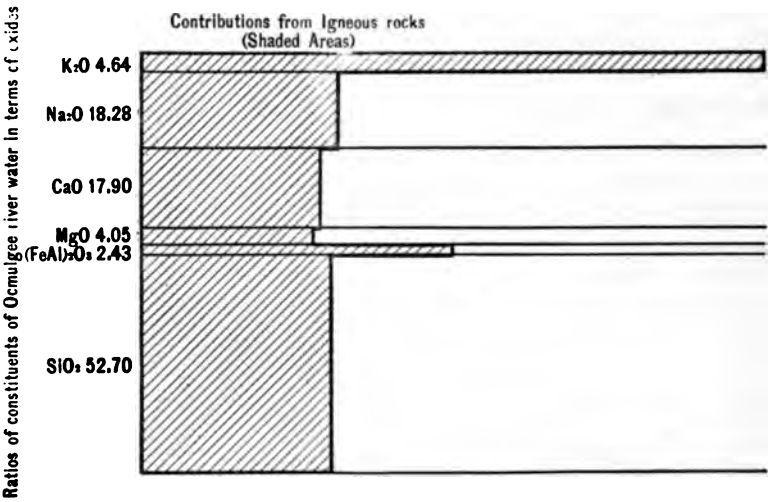


FIG. 22. Comparing graphically the losses by katamorphism of Georgia granites with the dissolved mineral content of the Ocmulgee river at Macon, Georgia.

the dissolved mineral matter of the Ocmulgee river [1] at Macon, Georgia, is expressed in terms of oxides. These analyses are compared graphically in Fig. 22 by the methods employed in

[1] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, p. 68.

Figs. 21 and 23. The shaded area represents contributions by weathering of granites and is in closer accord with the composition of river water than in either of the other two cases referred to.

Rivers Draining Areas of Sedimentary Rocks

Where rivers drain areas of sedimentary rocks containing limestone they are likely to show more of the lime and magnesium carbonates and soda, and less free silica and potash. This is also true even where limestone is not present in the sediments, because of easily soluble calcareous cements and layers in the fragmental sediments.

The Genesee river of New York drains an area consisting entirely of sedimentary rocks and the Ocmulgee river of Georgia drains an area entirely underlain by igneous and crystalline rocks. Analyses of the dissolved mineral content of these rivers are given in Table XVI for purposes of comparison. The Genesee river is higher in calcium and magnesium and lower in sodium, potassium, silica, and $(Fe,Al)_2O_3$ than the Ocmulgee river. The greatest differences are in calcium and magnesium.

TABLE XVI. COMPARISON OF DISSOLVED MINERAL CONTENT OF THE GENESEE RIVER, DRAINING SEDIMENTARY ROCKS, WITH THE OCMULGEE RIVER, DRAINING IGNEOUS ROCKS

	Genesee river at Rochester, New York [1]	Ocmulgee river at Macon, Georgia [2]
CO_3	37.94	21.06
SO_4	25.29	7.48
Cl	1.41	4.28
Ca	24.48	9.62
Mg	5.29	1.83
Na	2.59	10.23
K	1.35	2.90
SiO_2	.82	39.70
$(FeAl)_2O_3$	.83	1.83
	100.00	98.93
Salinity parts per million	170.	69

[1] Clarke, F. W., op. cit., p. 65.
[2] Clarke, F. W., op. cit., p. 68.

Rivers Draining Deeply Weathered Areas

The contributions from a given drainage basin are related to the speed and degree of rock decay as determined by climatic and topographic conditions. Under conditions of advanced decomposition, which have permitted the accumulation of a thick mantle of thoroughly decomposed rock material—usually in areas of moderate relief, abundant rainfall, high temperature, abundant vegetation, and slow erosion—the surface waters are obviously working over materials from which the easily soluble constituents have been abstracted. They are likely therefore to be of lower salinity, and to contain in solution relatively higher proportions of the more slowly soluble constituents left in the weathered mantle, such as potash, silica, and alumina.

The absolute amount of such substances dissolved under these conditions may not be greater than the amounts contributed earlier in weathering of an igneous rock, but the relative amounts are greater because the more readily soluble constituents have already been removed.

The Amazon system drains principally a country of low relief and deep and complete rock decay. The available analyses of the dissolved mineral matter in the Amazon tributaries [1] show an excess of silica over calcium, an excess of potassium over sodium, and a high percentage of $(Al,Fe)_2O_3$ (7 per cent to 13 per cent as compared with 2.75 per cent in Clarke's mean of river waters), with a low salinity averaging 45 parts per million. Apparently the least soluble constituents (those last to be removed in solution), silica, and the salts of potassium, iron, and alumina, are high because of the absence of the more soluble constituents, salts of calcium, magnesium, and sodium. Possibly the large amount of decaying vegetable matter contributes to the high potassium content of the waters. The silica may represent in part the results of lateritic decomposition of clays under conditions of extreme weathering. (See Chapter III, Part I.) In this connection it is perhaps significant that the rivers of India and certain tributaries of the Nile show a silica content much above the average for river waters.

[1] Clarke, op. cit., p. 83.

Rivers Draining Regions of Rapid Erosion

Where disintegration and erosion are rapid, as in certain arid regions of high relief, the waters are working constantly on comparatively fresh rock material, with the result that the dissolved mineral content may approximate more closely that calculated by comparing fresh and weathered phases of the rocks of the drainage basin. Also, under these conditions, there is often a concentration of the more highly soluble products of rock decay at the surface, with the result that these play an important part in the dissolved mineral content of the rivers. Weathering under arid conditions may result in accumulation of soluble salts in excess of the transportation facilities, resulting in high salinity of the few streams present. Higher salinity does not mean that more material is dissolved per unit of rock in these areas, but merely that there is less water per amount of liberated salts. This is illustrated by the high salinity of the rivers of the southwestern United States,[1] which have an average salinity of 1,994 parts per million as compared with 195 parts per million for the Mississippi river. Clarke [2] has estimated that the Mississippi river carries in solution annually 78 metric tons of material per square mile of drainage area, and that the Colorado river, which drains an arid country, removes in solution to the Gulf of California 59.6 metric tons per square mile per year. The greater flow of rivers of low salinity may thus more than balance the high salinity of rivers from arid regions.

Average River Waters

Where a river drains an area containing both igneous rocks and sediments the dissolved mineral content is intermediate between the two cases indicated. The solubility of limestone and dolomite formations is so much greater than that of other rocks, that the effect on the dissolved mineral content of rivers is out of proportion to their area in the drainage basin.

In Table XVII Clarke's [3] average of river waters, recalculated into terms of oxides, omitting the acid ions and minor constit-

[1] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, p. 77.

[2] Op. cit., pp. 71 and 77.

[3] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, p. 106 (column F).

uents, is given in comparison with the losses by katamorphism of average igneous rocks. (See p. 249.)

TABLE XVII. COMPARISON OF LOSSES BY WEATHERING OF AVERAGE IGNEOUS ROCK WITH THE DISSOLVED MINERAL CONTENT OF AVERAGE RIVER WATERS. (THE DATA ARE EXPRESSED IN TERMS OF OXIDES RECALCULATED TO TOTAL 100.)

	Losses by weathering of average igneous rock	Average of river waters
SiO_2	63.28	19.75
Al_2O_3, Fe_2O_3	2.71	4.66
MgO	5.53	9.64
CaO	13.36	48.30
Na_2O	8.85	13.20
K_2O	6.27	4.32
Totals	100.00	99.87

In the simplest terms the average river waters show relatively higher calcium, magnesium, and sodium, and lower silica and potassium, than the losses from igneous rocks by katamorphism. The most important cause of these differences is that the rivers derive their dissolved materials not only from igneous rocks but also from sediments. Sedimentary rocks yield material in solution by *solution of carbonates* more largely than by *decomposition of silicates*. Locally saline deposits yield solutes high in sodium chloride.

In Fig. 23 the average mineral content of river waters and the losses from the igneous rocks by katamorphism are compared graphically. The vertical dimension of the rectangle is divided to represent the ratio of constituents of average river waters in terms of oxides. The losses from igneous rocks by katamorphism are represented, in terms of oxides, by the shaded areas. It does not seem probable that all of the silica has come from igneous rocks, as the sedimentary rocks certainly yield silica to surface and underground waters. Clarke,[1] using the estimate by A. von Tillo, assumes that the land surface is comprised of approximately 75 per

[1] Clarke, F. W., The data of geochemistry: Bull. 491, U. S. Geol. Survey, 1911, p. 107.

cent sedimentary and 25 per cent igneous and crystalline rocks. On the assumption that these classes of rocks yield soluble salts in proportion to their areas, the relative contributions from igneous and sedimentary rocks have been represented in Fig. 23 by making the rectangle of such length that the shaded area represents one-fourth and the unshaded area three-fourths of the total area. This assumption is obviously open to criticism, and no emphasis is placed on the specific quantitative results. Qualitatively, however, the diagram indicates that the contributions from the sedimentary

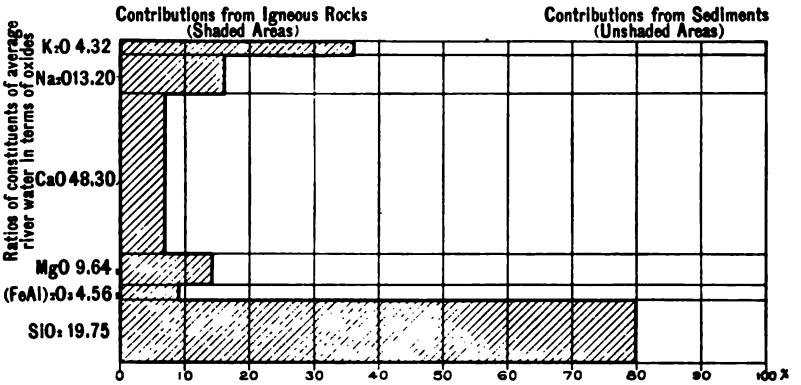


FIG. 23. Comparing graphically the losses by katamorphism of igneous rocks with the average dissolved mineral content of rivers.

rocks are relatively much higher in calcium, magnesium, and sodium, and lower in silicon and potassium, than the contributions from igneous rocks. It seems obvious that the *total* amount of sodium contributed by sedimentary rocks is much greater than that from igneous rocks, because of the greater quantity of sedimentary rocks on the erosion surface, although the percentage of soda to other salts contributed from sedimentary rocks is less than from igneous rocks. This is in part from undecomposed primary silicates in the sediments, but the large amount contributed, together with the high chlorine content of river waters, suggests that much of the sodium may be from sodium chloride included in the sediments in salt deposits, in connate sea waters, and in residually accumulated salt solutions found in deep waters (pp. 246–247) which ultimately reach the runoff by the lowering of the erosion plane.

If the sodium content of rivers represents in any considerable measure sodium once in the ocean and now on its *return* to the sea, a modification of estimates of the age of the ocean on the basis of sodium is obviously necessary.

"Solubility Ratios"

In a study of the solubilities of chemical constituents of rock,[1] C. H. Smyth, Jr., by comparing the materials carried in solution in rivers with the average composition of the rocks of the drainage area has computed the "relative solubilities" of the principal rock constituents. He finds the solubilities of the principal oxides to be in the following order, beginning with the most soluble—lime, soda, magnesia, potash, silica, alumina, and ferric oxide, and assigns to each of these a numerical figure for solubility, giving lime a solubility of 100. Smyth determined the ratio of solubility from a variety of data, and his results are summarized in the following table, giving lime a solubility of 100 in each instance.

TABLE XVIII. SUMMARY OF RELATIVE SOLUBILITIES FROM ARTICLE BY C. H. SMYTH, JR., JOUR. GEOL., VOL. 21, 1913, PP. 105 ET SEQ.

	Mean river compared with average surface rock	Mean river compared with average igneous rock	Mississippi river compared with average sediment	Ottawa river compared with average crust
	A	B	C	D
SiO_2	3.7	3.3	2.2	6.9
Al_2O_3, Fe_2O_3	2.6	2.0	.4	3.4
MgO	36.3	24.5	65.5	40.1
CaO	100.00	100.00	100.00	100.00
Na_2O	96.1	42.5	152.8	38.5
K_2O	11.9	10.4	12.5	13.4

The order of solubility as expressed by the losses of constituents in the weathering of igneous rocks may be read from the composite

[1] Smyth, C. H., Jr., The relative solubilities of the chemical constituents of rocks: Jour. Geol., vol. 21, 1913, pp. 105–120.

diagrams, Plates II and IV, showing respectively for acid and basic igneous rocks the losses and gains of constituents in weathering. These diagrams show the order to be as follows, beginning with the most soluble, CaO, Na_2O, MgO, K_2O, SiO_2, Fe, which order is essentially the same as the one obtained by Smyth.

DISSOLVED MINERAL CONTENT OF ENCLOSED BASINS AND SEAS

Waters become increasingly saline by evaporation in enclosed basins. The percentages of chloride and sulphate radicals increase and the percentage of carbon dioxide decreases. The relative importance of the acid radicals in fresh surface waters is $CO_2 > SO_4 > Cl$. In the sea this order is reversed. The bases also change in the following general proportions: in fresh water, calcium > sodium, sodium > magnesium, magnesium > potassium; in sea water, sodium > magnesium, magnesium > calcium, calcium > potassium. It is clear that the salt waters contain merely the residual of the substances which have come into solution.

Ocean water may be considered in a sense as a modified form of river water. This idea is expressed in the following approximate equation:

Ocean salts = possible primary ocean salts + contributions of rivers + atmospheric and volcanic contributions + marine solution of mineral matter — mineral matter removed from solution by precipitation and otherwise.

A comparison of the average dissolved mineral content of rivers with the average composition of ocean salts should indicate the relative changes in proportions of constituents. This problem admits of treatment similar to that used in comparing analyses of fresh and altered rock phases.

Considering the dissolved material of the average river water as an original phase and the average ocean salts as a resulting alteration product, the relative losses and gains have been expressed in a "straight-line" diagram in Fig. 24. For convenience each constituent is platted on a separate horizontal line. The horizontal scale represents for each constituent considered, in figures at the top of the diagram, the number of grams of ocean salts required to contain the amount of that constituent originally

present in 100 grams of average river water salts. The diagram shows the relative behavior of the several constituents. Relative to chlorine, all the constituents have been lost from solution. Relative to sodium, chlorine has been gained and all other constituents lost. If we assume that all of the sodium contributed to the ocean has remained in solution it follows that magnesium, potassium, calcium, iron, aluminum, and silica have been removed from solution in proportions increasing in the order given, that the sulphate and carbonate radicals have been lost from solution, and that chlorine has been introduced into the sea from other sources, possibly from volcanic emanations.

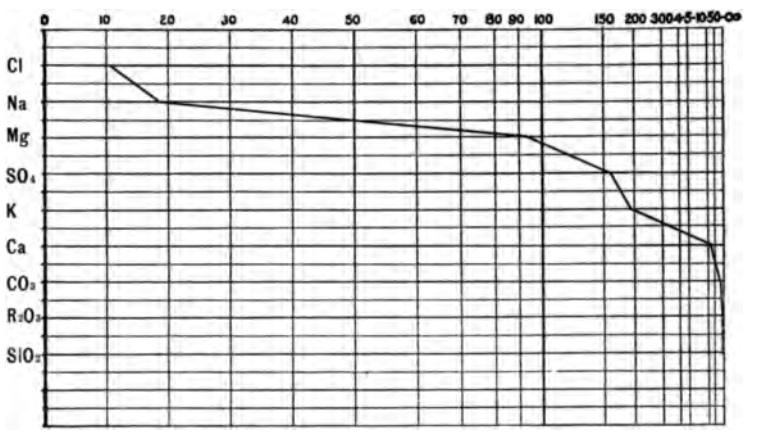


FIG. 24. Dissolved mineral matter of average river water compared with average of ocean salts.

Silica makes up nearly 12 per cent of the dissolved mineral matter of average river waters, and less than 1/100 of 1 per cent of the mineral matter of the ocean. From Fig. 24 it is evident that relative to sodium silica, has been lost from solution to a very large extent. This is also evidenced by the practical absence of silica from the ocean salts. In other words, a very large part of the silica contributed in solution to the ocean has been precipitated. Silica is precipitated to a large extent with finer clastics near the mouths of rivers and is also used in the shells and tests of organisms, the remains of which make up the vast deposits of siliceous oozes.

Alumina is carried to only a very limited extent by rivers and its percentage in the ocean salts is negligible.

Iron is carried in small amount by the rivers and its percentage in the ocean salts is extremely minute. Iron is known to be precipitated in marine sediments in the form of carbonate, sulphide, and silicate,—the first probably by the replacement of other carbonate, the second by reduction of the sulphate by organic materials, and the third largely as glauconite, through some process of replacement or chemical reaction.

Calcium constitutes slightly over 20 per cent of the dissolved mineral content of rivers and approximately 1.2 per cent of the ocean salts. Relative to sodium, calcium has been precipitated to a very large extent. The presence of calcium in appreciable amounts in the ocean salts shows that all of the calcium contributed to the ocean has not been precipitated. Calcium is removed from sea water principally to form the shells, tests, and skeletons of organisms. Chemical reactions precipitating calcium are also possible, and it is thought that calcium is precipitated from solution to a certain extent by the mechanical elimination of carbon dioxide, due to wave action. The cycle of carbonate deposition near the surface and resolution of calcic remains at depth tends to maintain the reserve of calcium in solution.

Magnesium constitutes 3.4 per cent of the dissolved mineral matter of the average river waters and 3.7 per cent of the ocean salts. These figures and the "straight-line" diagram, Fig. 24, indicate that magnesium has been contributed in excess of removal from solution, and that although magnesia has been lost from solution relative to sodium, a large reserve has accumulated in the ocean. Magnesia is used only to a very small extent by marine organisms, and its principal means of precipitation seems to be by reaction with previously precipitated calcium carbonate, in which it replaces part of the calcium. (See pp. 149–150.) Potassium makes up 2.12 per cent of the average river waters and 1.1 per cent of the ocean salts. Its presence in the ocean in this amount shows that more potassium has been contributed than has been precipitated. Relative to sodium, potassium has been lost from solution to a large extent. Potassium is removed from ocean water by marine plants and also by adsorption by the finer argillaceous sediments. Potas-

sium salts are present in some of the chemical sediments, principally in glauconite.

Sodium makes up 5.79 per cent of the dissolved material of average river waters and 30.59 per cent of the ocean salt. The lack of agencies for the precipitation of sodium from sea water has resulted in its accumulation to a very large amount. It has been calculated that the total salt in the ocean is complementary to an average thickness of 1.39 miles of sediments over the continental areas, on the assumption that the salts of the ocean have been derived from the destruction of preëxisting igneous and crystalline rocks by the ordinary processes of katamorphic redistribution. (See p. 73.) Since this calculated average thickness of sediments for the continents is not in excess of the actual amount of sediments as inferred from observation, it is not necessary to assume an original sodium content for the ocean. If the sodium content of the sea is the result of katamorphism of rocks, it seems probable that it has progressively increased through geologic time. If the rate of chemical denudation of land areas may be inferred from calculations of the sodium annually contributed to the ocean by rivers, the age of the earth may be estimated by dividing this figure by the total sodium content of the ocean, but there are wide variations in estimates of the present rate and in inferences as to changes of rate in the past. Also, as suggested on page 256, some sodium is deposited with the sediments and is again returned to the ocean, so that part of the sodium in rivers has already been in the ocean and should not be counted in estimating the age of the earth. How much correction of figures this factor involves is not known.

On the assumption that the chloride waters of deep mines and borings represent fossil or connate sea water, and that the salt water in a given horizon is of the same age as the rocks in which it occurs, Lane [1] concluded from comparison of these waters that the ratio of sodium to chlorine in the ocean has increased through geologic time. The assumptions used in reaching this conclusion may be open to question. (See pp. 246–247.)

The substances which are contributed to the ocean in excess of marine precipitation, principally chlorine, sodium, magnesium,

[1] Lane, A. C., The chemical evolution of the ocean: Bull. Geol. Soc. Am., vol. 17, 1906, p. 691.

sulphur, potassium, and calcium, represent to a certain extent materials split off or lost from the metamorphic cycle. This is true so far as they do not go back into rocks in the form of sediments or cements. The relative proportions of lime, soda, and potash, in the ocean, thus split off, are nearly those of blood plasma of vertebrates. As the result of extensive investigations of the inorganic composition of blood, MaCallum [1] concludes: "Enough, however, has been advanced here to make it extremely probable that the inorganic composition of blood plasma of vertebrates is an heirloom of life in the primeval ocean." This affords suggestive relationship between the by-products of the metamorphic cycle and the origin of life in ocean water.

[1] MaCallum, A. B., The inorganic composition of the blood in vertebrates and invertebrates, and its origin: Proc. Royal Soc. of London, vol. 82, ser. B, 1909–1910, p. 624.

CHAPTER III

THE METAMORPHIC CYCLE AS A BASIS FOR THE GENETIC CLASSIFICATION OF COMMERCIAL MINERAL PRODUCTS

The commercially important mineral products are largely segregated or altered by the processes of the metamorphic cycle. The study of these materials has yielded some of the best evidence of the nature of metamorphic processes. In one sense they may be regarded essentially as by-products of the metamorphic cycle. Practically all genetic classifications correlate mineral products to some extent with phases of the metamorphic cycle, but seldom is the order of the cycle closely followed, the tendency being to treat them as if they were a special class of substances developed by special processes rather than as incidents of general processes of metamorphism. Some advantage is to be gained by regarding mineral products in relation to the major metamorphic processes, in that this tends to bring out in better perspective the relations of ore depositing processes to metamorphic processes in general, and brings to bear on the problem a wide variety of observations and principles derived from a general study of metamorphism, which might not otherwise be so conveniently available. This treatment applies equally well to metallic and non-metallic products. It may be generalized for large groups of minerals or worked out in detail for individual mineral products without changing the perspective afforded by consideration of the metamorphic cycle as a whole.

The metamorphic cycle is ordinarily conceived to start with the igneous rock as the primary source of materials, follows the alteration of the igneous rocks into various end-products of katamorphism, follows these in turn through erosion and transportation, and then back to conditions of schists or gneisses or to their possible refusion as igneous rocks. At any point the cycle may be "short-circuited." Igneous agencies may bring about their pecul-

iar and special effects at any stage of the cycle, usually in the direction of accelerating and intensifying the processes characteristic of that phase. The cycle may be repeated many times. Ores may likewise be followed through the cycle.

Some illustrative examples of the classification of commercial minerals under the metamorphic cycle follow:

CLASSIFICATION OF IRON ORES

Next to alumina, iron is the most common and widely distributed metal of the earth and appears in every phase of the metamorphic cycle. It thus serves well as an illustration of the possibility of genetic classification under the metamorphic cycle.

1. Iron ore may occur as titaniferous magnetite formed as magmatic segregations in original igneous rocks. It may occur in veins and dikes deposited by hot waters as after-effects of igneous intrusion. It may be deposited from thermal solutions replacing rocks, particularly limestone, along igneous rock contacts; here also it may be regarded as an igneous after-effect. In all of these cases the primary source of the iron-bearing solutions seems probably to be the magma. Positive evidence is that the ores are formed at high temperatures, in association, both in time and place, with igneous rocks, and negative evidence is that it is difficult to account for them in any other way. Ores thus formed are hard and crystalline and are not easily altered by any of the processes of the metamorphic cycle, although they may undergo slight oxidation, hydration, and an enrichment by leaching of gangue materials.

2. Where iron is combined in the silicates of igneous rocks it may be concentrated by the processes of katamorphism which tend to break up the silicates. Under katamorphism some of the iron tends to be oxidized and hydrated and to remain as an insoluble residue. Where the rock is originally rich enough in iron, or the other constituents are relatively more soluble, the iron residue may be rich enough to be an iron ore. Such has been the origin of the Cuban limonites. If ores of this type become anamorphosed they are rendered crystalline, anhydrous, and magnetic, as the Cle Elum ores of Washington.

Exactly the same processes of weathering which develop a clay or a bauxite from granite or syenite develop iron ore from the Cuban peridotites or from the Lake Superior sedimentary iron

formations (see paragraph 4), the difference in results being due to differences in the primary rocks. The study of concentration or iron ores is therefore aided by an understanding of the principles of katamorphism as applied to other rocks.

3. During weathering part of the iron is taken into solution and transported and deposited as:

(a) Bog ores in the form of limonite;

(b) Extensive sedimentary formations, consisting of ferrous carbonate or ferrous silicate. Ordinarily such sediments are not sufficiently concentrated to be of commercial value. If the carbonate and silicate sedimentary ores are anamorphosed, amphibole magnetite formations result, which are seldom rich enough to be of commercial value. Important quantities of these are known in the Lake Superior country and other parts of the world.

(c) Extensive hydrated ferric oxide sediments like the Clinton ores. Anamorphic equivalents of the Clinton hematites are found in the less hydrous crystalline hematites of the Annapolis Valley of Nova Scotia. Brazilian hard hematites are also regarded as anamorphic equivalents of hydrated hematites of primary sedimentary deposition.

4. When the sedimentary iron formations of the carbonate and ferrous silicate types (b) are exposed to katamorphism, they are oxidized and hydrated and become concentrated, principally by the leaching of lime, silica, or other impurities, which leaves the ores as residuals. Such are the greater part of the Lake Superior hematites and the brown ores of the Appalachians. The anamorphism of these ores does not greatly change their composition, but tends to form crystalline, anhydrous hematites, with more or less magnetite, locally schistose, associated with heavy anhydrous silicates and quartz.

5. Iron in solution from the katamorphism of iron-bearing sediments may be distributed like the iron carried away from weathering igneous rocks. (See paragraph 3.) Ores may be transported, deposited and katamorphosed in successive cycles, leading to similar results. The most abundant and desirable ores are the result of katamorphic agencies of the second and later cycles.

6. The cycle may be short-circuited by the direct transfer of magmatic solutions from igneous rocks to the agencies of sedimentation, there to undergo successive cycles of katamorphism

and concentration. This is supposed to be the history of certain of the Lake Superior hematites.

CLASSIFICATION OF COPPER ORES

Turning to the copper ores, it appears that magmatic segregations of commercial value are rare, unless some of the primary ores of the "porphyry coppers" belong in this category. Most of the primary copper ores are developed as igneous after-effects, in veins or as replacements in igneous rocks or along igneous contacts, or as amygdaloidal fillings or replacements in surface flows and associated rocks, as in the native copper ores of Lake Superior. Where the primary ores are sulphides, katamorphism has greatly enriched them by developing oxide zones and zones of secondary sulphide enrichment. Erosion, surface transportation and deposition fail to develop important ore deposits of copper, although there are sedimentary beds carrying considerable quantities of copper, which in some places are rich enough for mining, as in the Mansfield slates of Germany. Further katamorphism of sedimentary copper formations tends to enrich them but yields no important copper deposits.

Copper is principally represented in stages of the cycle influenced by igneous agencies and by direct katamorphism in place, but so far as transportation and sorting become important, the segregation of copper ore diminishes. The most important ore bodies of iron, those formed from the katamorphism of sediments, do not have important correlatives in the copper ores.

CLASSIFICATION OF GOLD AND SILVER DEPOSITS

Gold and silver ores do not form important segregations in igneous masses. They occur principally as igneous after-effects in veins and in contact deposits, replacing wall rocks, and in replacements along contacts. Katamorphism tends to enrich the silver, principally by the leaching out of other constituents, but partly by carrying the silver downward in solution, replacing other constituents. It has very little effect on the gold. Because of its solubility silver plays less part in sediments than gold. Gold forms important placer deposits which may have been sorted by successive erosion cycles. Cementation and anamorphism may later recrystallize and indurate them. The

great intermediate stages of katamorphosed sedimentary ores, so important in the iron ores, are not represented in the gold and silver ores.

NON-METALLIC MINERALS

The metamorphic cycle is likewise a logical basis for the genetic classification of many non-metallic substances such as salt, phosphate, gypsum, borax, clays, bauxites, etc. Even coal and oil, although involving organic agencies in their development, which are not ordinarily considered in a discussion of the metamorphic cycle, are controlled largely by their relations to the cycle. Coal illustrates distinctive katamorphic and anamorphic processes.

GENERAL REMARKS

This manner of treatment of mineral products has its practical application in the study and teaching of economic geology. There is a growing tendency to get away from the notion that economic geology is a highly specialized phase of geology with its own body of principles, and to regard economic geology as merely the application of the broadest geologic knowledge to particular mineral groups that happen to be of commercial value. The use of some broad scheme of classification which applies not only to ores but to all inorganic substances, tends to promote this desirable tendency. It is perhaps not out of place to remark that the writers have used this method in teaching and study of economic geology for some years and find it workable. The particular method we have used is to follow each commercial group of minerals through the different phases of the metamorphic cycle, indicating where in the metamorphic cycle each of its important deposits should be classified. The repetition of this method for each of the mineral products amounts to a reiteration of the principles of metamorphism which tends to fix the major principles without getting away from the essential basis of inductive field study. To answer the question why certain mineral products are important in certain phases of the metamorphic cycle, and others are confined to different phases, requires a consideration of metamorphic relationships which are helpful to students in acquiring a ground work of economic geology.

CHAPTER IV

NET RESULTS OF THE METAMORPHIC CYCLE

IS THE METAMORPHIC CYCLE CLOSED?

Anamorphism tends to undo the work of katamorphism and may even bring katamorphic products back to an igneous condition. In this major cycle there are minor cycles which are clearly closed. For instance, consolidated sediments may again be distributed by processes of erosion, and again consolidated to their original conditions. Crystalline schists may run through the katamorphic phase of the cycle and back to the same condition. Yet within our range of observation there seems to be evidence that igneous rocks are not reproduced to the same extent that they are destroyed, that the cycle as a whole is not closed, that the result is an accumulation of by-products of the metamorphic cycle.

An attractive and plausible hypothesis, with a considerable number of supporters, is that the rise of temperature with depth or dynamic movement may liquify or fuse rock, and that magmas thus formed rise toward the surface and make way for themselves by a process of assimilation of the rocks of the earth's shell, aided by the mechanical stoping of the overlying rocks, allowing blocks to be ingulfed in the magma and then fused. This is the hypothesis of "subcrustal fusion" or "assimilation." This would mean a closing of the cycle so far as the fused rocks are metamorphic products. This conception has arisen principally from the difficulty of accounting otherwise for the space occupied by enormous plutonic intrusives, as for instance the Laurentian batholiths in the pre-Cambrian shield of North America, and the supposed absence of original basal portions of sedimentary strata, or of the basements upon which these strata were deposited. The possibility, or even the probability, of this having occurred on a large scale cannot be disproved, for, in the nature of the case, much of the evidence is destroyed by the process. After a rock is fused it is an igneous

rock. What its earlier history may have been is difficult to trace. It is equally difficult to prove the importance of this process. Our own observations on the old rocks of the pre-Cambrian lead to conclusions similar to those of Iddings,[1] who says:

"The overwhelming evidence of innumerable exposures of contacts between intruded rock masses and wall rocks of different kinds shows that, in the greater part of the lithosphere exposed to view, the blending by melting and solution of wall rock and molten magma has not taken place to any considerable extent. Dikes of granite traverse beds of sandstone, limestone, and shale without changes in the chemical composition of the granite at the contacts with these different bodies, such as should be expected, if they had been dissolved by the granite. From this we may conclude that in such parts of the lithosphere the intruded magmas were not hot enough to produce the effects claimed by the hypothesis mentioned."

Also Harker [2] says, with reference to the assimilation hypothesis:

"The difficulties—mechanical, thermal, and chemical—involved in such a theory have been sufficiently set forth by Brögger and others. The geological evidence seems to prove very clearly that these large igneous rock-bodies, like the smaller ones, are essentially *intrusive* as regards their relation to the environing rocks."

That depth alone has not been sufficient to cause fusion within the range of observation is shown by the fact that some of the very oldest pre-Cambrian rocks in the vicinity of great batholithic intrusions have not been fused but still remain as recognizable sediments or surface flows. It has been argued that the base of the sedimentary succession in the pre-Cambrian of Canada has been buried so deeply that it has been fused and is now represented by granite batholiths of Laurentian age. Yet in many parts of the pre-Cambrian, even in areas which have been cited as illustrations of the truth of this hypothesis, the original bases of the sediments have been found without evidence of fusion, and the batholiths prove to be intrusives which have worked their way above this basement. Whatever the *a priori* probability of fusion, the evidence cited from specific localities has been usually open to doubt.

[1] Iddings, J. P., Problem of volcanism, pp. 112–113, New Haven, 1914.
[2] Harker, Alfred, Natural history of igneous rocks, p. 83, New York, 1909.

Even if a batholith could be proved to be the fused lower part of the geologic column, it is to be remembered that these intrusions are on a small scale as compared with the earth as a whole, and do not furnish sufficient justification for the conclusion that the lower part of the geologic column has been generally fused by increased temperatures and pressures at considerable depth, and hence that the metamorphic cycle is in general closed.

Another line of evidence bearing on subcrustal fusion is furnished by a comparison of the amount of salt in the ocean with the thickness of sediments it is supposed to represent. It has been calculated (see p. 73), that the amount of salt in the ocean corresponds to an average thickness of sediments of 7,100 feet over the continental areas. Estimates of the actual average thickness of sediments, based on measurement of sections and other inferences, give results not widely different. In other words, the sodium of the ocean is in proper proportion to existing sediments, assuming both to have been derived by katamorphic redistribution of igneous rocks. If large quantities of sediments had been fused there should be a corresponding excess of sodium in the sea, over the amount comparable to existing sediments.

Observation seems to disclose gradual accumulation of sediments and their equivalent schists, since the earliest recognizable pre-Cambrian horizons. Some small parts have been converted to igneous rocks around igneous intrusions. Whether beneath the lowest rocks accessible to observation sediments have been more largely converted into igneous rocks, we have no means of knowing. It may be that the depth of the geologic column open to observation is too small a part of the geologic record from which to draw our inferences, but it is all we have as a basis of observation.

From the foregoing considerations we conclude that whatever the *a priori* probability of the closing of the metamorphic cycle on any large scale, adequate evidence of it is lacking, and that such evidence as there is points rather toward the incompleteness of the cycle, with a resulting accumulation of by-products of the cycle, consisting of sediments, schists, and salts of the ocean.

In terms of minerals this means, in general, the loss of feldspars, of certain ferromagnesian constituents like augite, and increase in the carbonates, oxides, hydrates, and in platy minerals of the hornblende, chlorite, and mica type.

SELECTIVE REDISTRIBUTION OF EARTH MATERIALS RESULTING FROM THE METAMORPHIC CYCLE

Whether or not the metamorphic cycle is closed on a large scale, it is clear that in its different phases there are selective processes at work which tend to accelerate the progress of some substances and retard the progress of others through the cycle, with resulting effects on the nature and proportions of the different classes of metamorphic products in a given phase of the cycle. Some of these effects stand out clearly, others may be merely suspected or inferred from the nature of the processes and the meager quantitative data available. This phase of study of metamorphic rocks offers a most attractive opportunity for further development as our knowledge of metamorphic processes becomes more clearly defined and based on quantitative observations.

Perhaps the must clearly recognized and defined result of the metamorphic cycle is the segregation of sodium in the sea, some features of which have already been described. There is no evidence that sodium thus segregated gets back to any extent into the anamorphic phase of the cycle or into igneous rocks resulting from fusion of sediments. So far as connate salt waters exist in sediments there may be opportunity for the recombination of soda in new minerals during anamorphism, but neither the existence of this water nor its combination with the minerals has been proved on any considerable scale. So generally is it accepted that the sodium in the sea does not get back into the rocks, that its amount is used as a measure of the chemical denudation which rocks have undergone by katamorphism.

As we follow the anamorphic changes of all kinds of rock, the fact stands out conspicuously that excess of lime is expelled by anamorphism. The schists and gneisses characteristic of anamorphism contain a minumum of lime-carbonate, or even of lime-silicate. To only a very limited extent lime enters into the constitution of platy or columnar minerals adapted to the anamorphic phase. In contact metamorphism also lime is eliminated. This expulsion of lime in anamorphism suggests a concentration of lime near the surface.

This may partially explain the apparent fact (see pp. 67–68) that the actual proportion of limestone to other sediments, as

measured in sections, seems to be larger than that calculated on the basis of comparison of analyses of sediments and primary igneous rocks.

Another possible cause of segregation of lime in sediments is the fact that basic rocks, high in lime, are, on the whole, much more readily attacked by the acid agencies of katamorphism. At any given stage of erosion relatively more of the completely decomposed products of the basic rocks, high in lime, may have gone into sediments, than the decomposed products from the weathering of acid rocks, which more largely yield by disintegration. The relatively acid and undecomposed character of many terrestrial deposits, known to be formed by disintegration, is evidence along the same lines.

So far as this tendency for segregation of lime at the surface is a fact it would tend to neutralize the so-called "acid shell" of sediments.

It does not follow from the above that no limestone has remained in the zone of anamorphism, for it is clear from the existence of marbles in areas of schists and gneisses that anamorphism has not succeeded in undoing all that katamorphism has accomplished in segregating lime.

Dolomites seem to be more abundant in older and more highly metamorphosed geological terranes than in later ones. As the detailed changes of the metamorphic cycle are followed magnesia is found to persist through the cycle to a much larger extent than lime, as attested by actual analyses of related series of specimens. The recurrence of cycles should lead to an increased percentage of lime over magnesia in the great carbonate formations. That limestones are in larger proportion to dolomite in later as compared with earlier formations seems to be a pretty well-established fact.[1] If it is true, it supports our inference of the progressive accumulation of limestone near the earth's surface.

Among the fragmental sediments the ones which most readily yield to rock flowage are shales, both by reason of mechanical weakness and their chemical and mineralogical characteristics. It is possible also that their exceptionally large content of alkalies, as compared with the other fragmental sediments, acts as a flux

[1] Daly, R. A., First calcareous fossils and the evolution of the limestones: Bull. Geol. Soc. Am., vol. 20, 1909, pp. 163–167.

which makes them more easily fusible, but the fusion of slates on a large scale has not been proved, whatever its probability. (See pp. 117–119.)

Quartzose sediments, on the other hand, when they go into the zone of anamorphism become hard and crystalline, and to only a slight extent schistose or fused. Their highly acid character usually remains. Anamorphism has not been sufficient to recombine these segregated products to produce schists on a large scale.

This difference in behavior of the shales and quartz sand formations under anamorphism should tend toward an increase in the recognizable sand formations as compared with the shales within our zone of observation. It has already been noted that the actual proportion of sands to other sediments, as measured in sections, is larger than that theoretically required on the basis of comparison of the chemical composition of sediments and that of igneous rocks, and as a partial reason for this it has been suggested that the sands include a considerable number of undecomposed materials derived principally by disintegration, in other words that some sands contain the constituents both of limestones and shales. (See pp. 67–68.) So far as sands are residually accumulated by their failure to enter into the schists or igneous rocks, this would tend further to explain their larger preponderance in actual sections.

It has been suggested (p. 68) that repeated cycles of katamorphic redistribution of sediments may have resulted in a progressive cleaning up of the sediments by separation of quartz and carbonates from the shales, thus increasing the proportions of sands and carbonates.

ENERGY CHANGES INVOLVED IN THE METAMORPHIC CYCLE

Much remains to be learned about the energy changes in common rock alterations, but enough is quantitatively known to enable us to state, with some confidence, that the changes of the metamorphic cycle result in a permanent net loss of energy from the system in the form of heat. It may be assumed from this and from *a priori* reasoning that it is this running down of energy which keeps the cycle going. In the igneous rocks energy is locked up in the molecular combination of minerals which under katamorphism

is liberated as heat and much of it permanently lost. As this change involves expansion, it is constantly opposed by gravity, while in the anamorphic phase of the cycle gravity controls, leading to a diminution of volume. We may suppose that at any given depth some sort of energy equilibrium exists between pressure, heat, and the energy locked up in molecular combinations. So far as a rock is not adapted to this state of equilibrium, changes will go on. A granite formed under conditions of high temperature and high pressure when brought to the surface finds itself under conditions of low temperature and low pressure. The expansive forces of the chemical reactions with the surface elements or expansive forces mechanically locked up in crystalline structure are but feebly resisted by gravity. The rock adapts itself to the new conditions of equilibrium. Unconsolidated sediments, when buried beneath the surface, reach new conditions of temperature and pressure, and readjustment is required to establish a condition of energy equilibrium. The constituents of the sediments are forced together by gravity, heat is gained, and energy is locked up in the molecular combinations, only to be lost again as the rock later comes under katamorphic conditions. The existence of large masses of anamorphic products, which have not reached the extreme stage of igneous rocks by fusion, implies that so far as these products have been accumulated energy has not been stored up in the anamorphic phase of the metamorphic cycle to the same extent to which it has been lost by katamorphism.

If heat is permanently lost by these cyclic changes and otherwise, thereby lowering the temperature gradient, while gravity remains constant, it may be supposed that the energy available for rock change at a given depth becomes less. In order that the rock may fully complete its cycle by storing up energy in one phase equal to that lost in another it must go to a point deeper within the earth, where increased effect of pressure and heat can make up for the loss of heat during the cycle. The running down of energy therefore implies the permanent accumulation of metamorphic products instead of reproducing igneous rocks on the original scale.

CONCLUSION

In summary, then, the metamorphic cycle may be regarded as indicative of a great pulsational alteration of the earth's surface, kept going through the running down of energy and its escape from the earth, the cycle being an expression and vehicle of this running down of energy. This cycle involves reversals of processes which keep the rock materials within certain limits of mineral condition and distribution, but these reversals are not quite compensating, with the result that there is a residual accumulation and redistribution of certain substances such as sediments, schists, and gneisses, and salts of the sea. Some of the accumulations are definitely known, some are inherently probable, some are merely suspected.

If there is truth in this general conception of the net results of metamorphic cycles, the study of metamorphism may become somewhat prognostic if we use it to direct our search for some of these larger cumulative results which are now obvious in detailed work. The discernment of these results will require a much wider range of study and quantitative observation than we have before attempted.

The acceptance of this view of the significance of the metamorphic cycle involves perhaps a slight modification of the prevalent interpretation of Hutton's great law of uniformitarianism, that the present is the key to the past. In metamorphism the same series of processes are operating to-day that have operated in the past, but they have slowly made over the rock types and redistributed them vertically and horizontally, with the result that while the same processes act to-day as have acted in the past, they are to-day operating, we infer, on different proportions and distributions of substances, with consequently differing emphasis and producing slightly different results.

PART IV.

LABORATORY WORK IN METAMORPHISM

INTRODUCTION

Laboratory study of metamorphism may include: (a) investigation of metamorphic processes and principles as inferred from field and laboratory study of metamorphic rocks; (b) work of a synthetic nature, in the attempt to reproduce natural conditions and to study the changes in rock material under controlled conditions of temperature, pressure, and time.

This discussion deals principally with the first of these two main divisions of laboratory work. The synthetic study belongs more to the physical chemist than to the geologist, and for the most part requires facilities not generally available to geologists. Important work of this type is being done at the Geophysical Laboratory of the Carnegie Institution at Washington. The difficulty of imitating natural conditions in the laboratory, the extreme complexity of the conditions and materials, and the impossibility of reproducing the time conditions, make it imperative that conclusions based on this type of laboratory work be carefully checked with facts observed in nature, in the field and in the laboratory. Likewise the inductive study of natural occurrences gains valuable aid from the synthetic work in the laboratory.

In the following discussion emphasis is laid on the study of metamorphic rocks as they occur in nature, not because this kind of laboratory work is necessarily more important than the synthetic study by the physical chemist, but because it naturally precedes the synthetic work in defining the field and conditions, and because it requires so little in the way of special equipment that it is within the range of most geologists, which cannot be said of the synthetic laboratory work. Specimens, samples, thin sections, descriptive and analytical data are available to some extent to all students of metamorphism. Laboratory work with such materials involves, among other things, macroscopic and microscopic study, measurement of physical properties, such as density and porosity, comparison of chemical analyses of successive stages in metamorphic changes in order to deter-

mine relative or absolute gains and losses, calculation of minerals from chemical analyses, and calculation of volume changes, both for individual mineral alterations and for alterations of rocks.

The purpose of laboratory work is to arrive ultimately at an understanding of processes of metamorphism and their significance. Too often the work has stopped with the accumulation of more or less uncorrelated data and descriptions. Throughout the following discussion the ultimate purpose of the methods described should be kept clearly in mind.

Careful discrimination between results of qualitative and quantitative study is essential. The data thus far available in metamorphism are very inadequate to a full treatment of the subject from a quantitative standpoint. Qualitatively the major steps in the processes of the metamorphic cycle are known, and these are confirmed in essential particulars by quantitative data, but in the study of almost any phase of the metamorphic cycle there is some lack of precise information. Chemical analyses are especially necessary to an understanding of metamorphic changes. In running over the list of analyses available we find rather surprising gaps in the information. In the plates of this volume (Pls. II, IV, V, VI, VIII, X, XI, XII, XIII, XIV, XV) we have collected and compiled chemical analyses bearing on the chemical and mineralogical changes involved in rock alterations of the principal phases of the metamorphic cycle. Some of the analyses hitherto unpublished have been made from our own investigations. The greater number are from published sources. While the best known sources have been carefully searched for analyses we make no claim that the list is complete. There are undoubtedly analyses which have not come to our notice. Some published analyses have been omitted because of doubt of their validity or application to the particular type of metamorphism they purport to represent. Others have been included which are not entirely satisfactory in this regard, but are the best available. There are undoubtedly many analyses included in these based on single hand specimens which do not properly represent the alterations which have occurred. Other analyses are included which are not complete, particularly as to ferric and ferrous iron. In all we have represented 111 sets of analyses representing metamorphic changes.

The preponderance of analyses showing weathering and hydrothermal alterations of igneous rocks and contact alterations of limestones shows how quantitative studies have been concentrated on apparently few phases of metamorphism, either because of economic importance or of such general distribution as to be open to study by many investigators.

In the following table the number of sets of comparable analyses known to the writers, representing metamorphic alterations, are given. These figures bring out the fact that several important phases of metamorphism have been given little or no quantitative study.

TABLE XIX. SHOWING THE NUMBER OF SETS OF AVAILABLE ANALYSES REPRESENTING METAMORPHIC CHANGES

	Weathering	Hydrothermal	Dynamic	Contact
Acid igneous...	16 sets	17 sets	1 set	[1]
Basic igneous ..	17 "	10 "	4 sets	[1]
Shales and slates	3 "	[2]	none	16 sets
Sandstone.....	2 "	none	3 sets	3 "
Limestone and dolomite	4 "	[2]	1 set	15 "

One of the most common conceptions of metamorphism is that certain gneisses are derived from the dynamic metamorphism of granites. Such cases have been often described. Many analyses of gneisses appear in the literature in comparison with granites, but so far as we have been able to ascertain, there have been presented no authentic sets of analyses of an unaltered granite and of a derived gneiss which can be accepted as truly representative of this alteration.

The dynamic metamorphism of rocks in general is so inadequately based on quantitative data that it is actually impossible today to tell what all of the significant chemical changes are. There seems to be qualitative evidence and quantitative evidence, in the few analyses available, that important changes go on during this process, as in all other phases of the metamorphic cycle, but

[1] Included with hydrothermal alterations.
[2] Included with contact alterations.

discussion of the subject cannot be satisfactory until many more analyses are available.

With all the careful study and discussion of limestone contacts in connection with ore deposits, quantitative chemical data are not yet available as to the composition of the contact zones on a large scale as compared with the unaltered rocks. There have been many analyses of individual hand specimens, but no proper samplings in the manner in which, for instance, an ore deposit would be sampled.

In the discussion of alterations of ore deposits, particularly the weathering and secondary enrichment, there is a curious lack of adequate quantitative data. One may look through a considerable number of the most satisfactory and detailed reports on mining districts, in which a large amount of attention is given to secondary concentration, without finding adequate large scale, quantitative determinations of the compositions of the several zones, as determined by mine runs through long periods.

An interesting gap in quantitative metamorphic data is in the cementation of sediments. Cementation may begin from the moment the sediments are laid down, and some little evidence is accumulating that the changes in chemical composition are continuous from that moment until the rocks are thoroughly cemented, but these changes have not been followed through in detail with reference to their locus.

Some of the alterations not adequately represented by analyses in the above table have been so often, and so well described, that the absence of analyses from these tables may be a surprise to the reader, and may raise the question whether available analyses have not been overlooked. We have studied many analyses purporting to cover these alterations but have not found that they can be used in strictly quantitative comparison, either because of lack of proper sampling, or more generally, because the fresh and altered phases are not strictly comparable. Many analyses of altered phases are given. Where the analyses of the fresh rocks, or unaltered rocks, are offered in comparison it appears that they are usually from rocks which have not been satisfactorily shown to be originally of the same composition. Samples may be taken a half mile or more away without adequate care to get the same beds or types. Study has been largely devoted to hand

specimens, often collected at random, and the analysis has frequently followed more or less as an afterthought. The greater number of analyses have been made for petrographic identification and not as interpretative of metamorphism. Still another difficulty is that so many analyses are incomplete in regard to principal oxides.

The collection of proper samples to illustrate metamorphic changes is not an easy matter. Only in a few favored localities are exposures sufficiently continuous, and the conditions sufficiently clear to make it possible to collect samples which are properly representative of the changes.

While much is known qualitatively about certain phases of metamorphism, which are not adequately represented by quantitative data, it seems to us certain that the development of quantitative information about these types of metamorphism may change our conception of these processes, and of the important factors in these processes, in important particulars.

One of our purposes in publishing the available data in present form is to call attention to these remarkable gaps in our quantitative information, in the hope that certain additional analyses may be called to our attention, and also in the hope that geologists in the field will take advantage of opportunities to make quantitative determinations to fill these gaps.

In the absence of related series of analyses much can be done in getting a broad view of the chemical changes in metamorphism, by comparing average analyses of types of rocks known to be successively developed by metamorphic processes, without regard to location, for instance average shales with average slates, or schists with possible sources.

CHAPTER I

METHODS OF LABORATORY STUDY OF METAMORPHIC ROCKS

MACROSCOPIC AND MICROSCOPIC STUDY OF SAMPLES AND SPECIMENS

A knowledge of petrology and petrography is prerequisite to this work. Thin sections and specimens should be studied from the standpoint of *metamorphic changes and processes*, rather than simple description of rocks and minerals. This work should be correlated with the chemical changes when analyses are available. In studying a mineral alteration it is desirable to write equations for the change [1] and to calculate the losses and gains of constituents and the density and volume changes involved. Some idea of the quantitative range of metamorphic changes studied may be obtained by measurement of proportions of minerals by the method of areal measurement under the microscope.[2]

In certain metamorphic relations the nature and abundance of certain heavy accessory minerals become important.[3] These minerals, because of very small size and limited amounts, cannot be studied to advantage either in hand specimens or in thin sections. However, their high specific gravity and hardness make it possible to concentrate them mechanically. The rock samples or specimens should be crushed to pass through a sieve slightly coarser than the grains to be examined. This crushing can be done on a small scale in a diamond mortar, or on a bucking board. If an ore dressing laboratory is available, it is desirable to crush the mate-

[1] Van Hise, C. R., A treatise on metamorphism: Mon. 47, U. S. Geol. Survey, 1904, Table C, p. 375.

[2] For discussion of methods see Johannsen, A., Manual of petrographic methods, New York, 1914, p. 290.

[3] Derby, O. A., On the separation and study of the heavy accessories of rocks: Proc. Rochester Acad. Sci., vol. 1, 1891, p. 202.

rial by means of the steel rolls. The crushed material should be sized by means of a set of sieves and the lighter material removed by washing or panning. The minerals in the concentrate may then be further separated by means of heavy solutions, such as the Thoulet solution. They can then be examined under the microscope.

A binocular microscope, such as is commonly used in biological work, possesses many advantages over the monocular microscope in working with rock fragments and specimens in ordinary light.

A microscope arranged for examination of polished surfaces by reflected light is very useful, especially in studying opaque minerals and ores.

MEASUREMENT OF SPECIFIC GRAVITY AND POROSITY

The specific gravity of rock specimens and minerals is determined by the common water displacement methods or in the powdered form by means of the picnometer. The picnometer determination of the specific gravity of finely-crushed or powdered rock material has certain advantages over the determination of specific gravity of hand specimens, as in the latter it is difficult to insure complete saturation when weighing under water.

Porosity or percentage of volume occupied by voids can be determined in specimens of coherent material by determining the weight in grams of water required to completely fill the pore space. The volume of a rock *exclusive of voids* is expressed in cubic centimeters by dividing its weight in grams by its mineral specific gravity.[1] The volume of the voids is expressed in cubic centimeters by the weight in grams of the water required to fill these voids, in other words, by the difference in weight between the dry and saturated rock. The percentage of volume occupied by voids is expressed by the following equation $P = \frac{W \times 100}{W + \frac{R}{G}}$ in which P equals percentage of pore space, W equals weight in grams of absorbed water, R equals weight of dry rock in grams and G equals mineral specific gravity of rock.

[1] By *mineral specific gravity* of a rock is meant the specific gravity of the mineral content of the rock, excluding the modifying effects of pore space and moisture. The term *rock specific gravity* is used with reference to the specific gravity of the rock in its natural state.

In very dense rocks, having a small percentage of voids, the porosity may be determined by making determinations of the specific gravity of the rock specimen and of the finely-ground rock powder, the porosity being calculated from the difference in these two figures. In determining porosity by this method the specimen should be dried thoroughly and then weighed in water as quickly as possible to avoid absorption of water. If the porosity of the specimen is very slight the water is absorbed so slowly during weighing as to cause only a small error. If the porosity is large enough to admit of rather rapid absorption, it is necessary to coat the specimen with paraffine or a similar substance to prevent the absorption of water. This coating in itself introduces a certain amount of error and is a rather questionable procedure. A better method for determining porosity is by measuring the amount of water absorbed and in the case of dense specimens to obtain accuracy by careful weighing and by prolonged boiling of the specimen under partial vacuum, to insure thorough saturation before weighing under water. In all cases it is advisable to determine the weight of the water-saturated specimen and the weight of the saturated specimen under water before determining the weight of the dried specimen, as the undried specimen contains water in the finest pores, which it is difficult to replace subsequent to drying. If the moisture of the saturation is determined in terms of percentage weight of saturated rock, (moisture of saturation) the porosity may be read directly from the diagram, Plate XVI.

The porosity of unconsolidated or incoherent materials, such as residual clay or soil, or altered rock too soft and porous to be collected in the form of a solid hand specimen, may only be determined in the field, as follows: Thoroughly saturate a portion of the material *in place*, then remove a suitable sample with a shovel or trowel and place it immediately in a tightly-covered receptacle. From this sample the moisture of saturation should be determined, and since the specimen was saturated in place, this moisture will represent the amount of water required to fill the pore space, and is, consequently, a measure of the natural porosity of the sample. Specific gravity may be determined from a portion of the sample by means of a picnometer and the porosity read from the diagram, Plate XVI.

LATE XVI

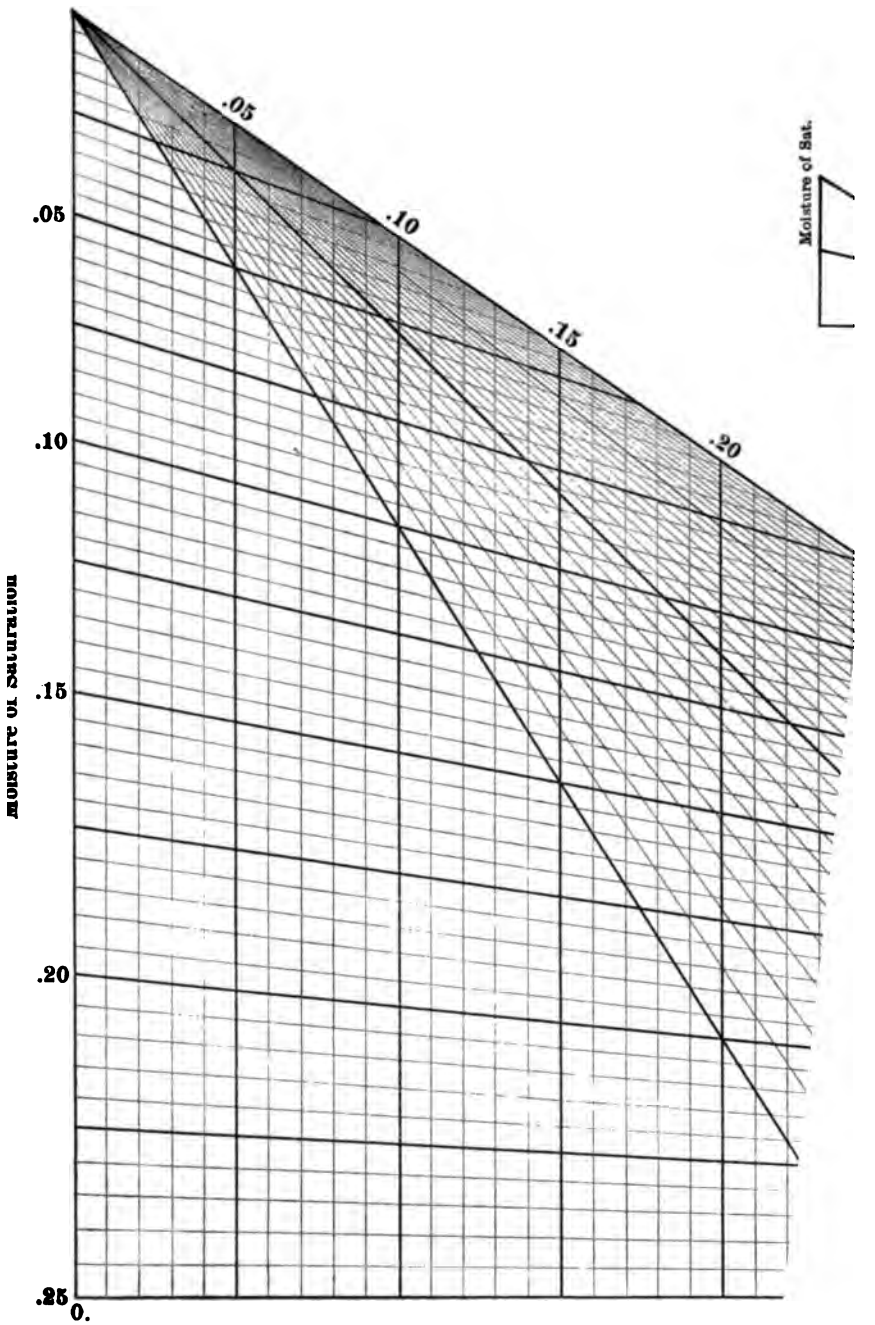
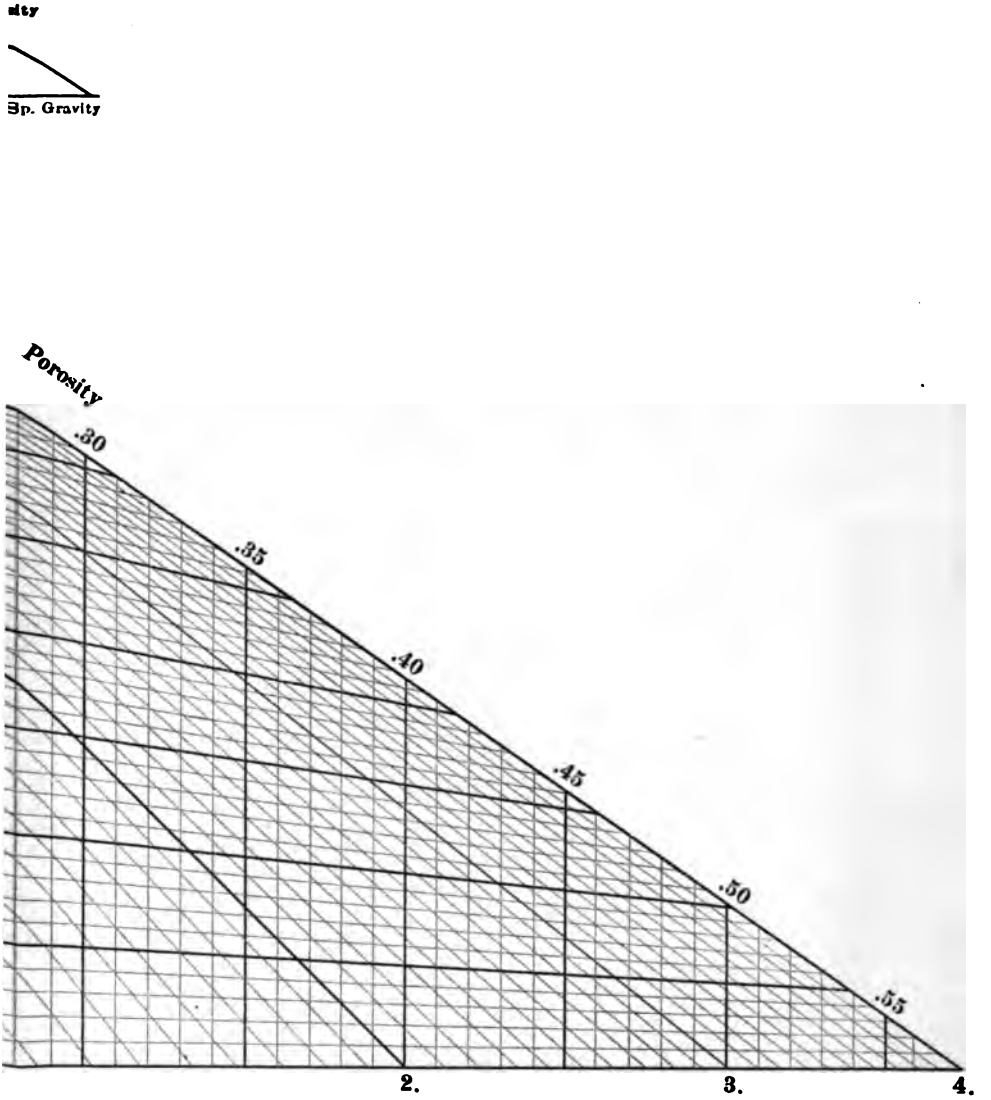


Diagram for the calculation of poros



fic Gravity

of saturation and specific gravity of rock materials.

COMPARISON OF ANALYSES TO DETERMINE RELATIVE AND ABSOLUTE GAINS AND LOSSES OF CONSTITUENTS

Metamorphic changes in rock in general involve changes in chemical composition due to losses and gains of constituents. These losses and gains can best be studied and determined quantitatively by comparison of analyses representing the rock in different stages of alteration. Reliability of results based on chemical analyses of rocks obviously requires proper field investigation of the rocks to insure that the samples analyzed are really comparable, and requires careful sampling of the rocks to be analyzed.

Chemical analyses represent the relative amounts of the various constituents in terms of weight. Direct comparison of analyses representing different stages in the metamorphism of a rock, can be made only when there has been no change in weight due to the alteration. If 100 grams of fresh rock produce 100 grams of altered rock the losses and gains may be found by direct comparison of the analyses of the fresh and altered rock. When, due to losses or gains of constituents, 100 grams of fresh rock have produced *more* or *less* than 100 grams of altered rock, obviously direct comparison of chemical analyses of the two phases cannot be made. In that case it is necessary to reduce the analyses to a common basis for comparison.

In alterations which have proceeded without change in volume, weight relations between unit volumes of the fresh rock and the altered phases may be calculated, if the mineral specific gravities and the porosities of the two phases are known. Ransome [1] has calculated gains and losses, on the assumption of constant volume, in hydrothermal metamorphism of rocks from Goldfield, Nevada, and from Breckenridge, Colorado.

When weight or volume relations between the two phases are not known, it is possible in some cases to select a constituent which has remained nearly constant in absolute amount during the alteration. With this as a basis the analyses may be recalculated to a comparable basis. (See calculation on basis of constant alumina in weathered Georgia granite, pp. 7–10.)

[1] Ransome, F. L., The geology and ore deposits of Goldfield, Nevada: Prof. Paper 66, U. S. Geol. Survey, 1909, p. 181; Geology and ore deposits of the Breckenridge district, Colorado: Prof. Paper 75 U S Geol. Survey, 1911, p. 96.

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[1] Ransome, F. L., The geology and ore deposits of Goldfield, Nevada: Prof. Paper 66, U. S. Geol. Survey, 1909, p. 181; Geology and ore deposits of the Breckenridge district, Colorado: Prof. Paper 75 U S Geol. Survey, 1911, p. 96.

" Straight-Line " Representation of Gains and Losses of Constituents

In cases in which no single constituent has remained constant in amount and in which weight or volume relations are not known the *relative* gains and losses of constituents may be expressed by the quotients obtained by dividing the percentage of each constituent in the unaltered phase by its percentage in the altered phase. These quotients multiplied by 100 and platted to scale on a line from a single zero point constitute a "straight-line" diagram. They indicate for each constituent the number of

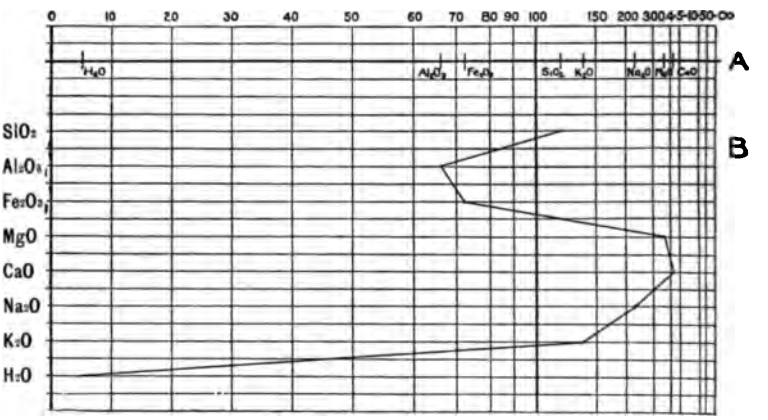

FIG. 25. Illustrating two methods of platting a "straight-line" diagram.

grams of altered rock required to furnish the amount of that constituent originally present in 100 grams of unaltered rock. The values for the several constituents may all be platted on the same line or each may be represented on a separate line. These two methods are shown respectively at A and B in Fig. 25, representing the gains and losses of a Georgia granite by weathering from analyses in Table II, p. 8.

Since these values may vary from zero (for a constituent absent in the fresh rock and present in the altered) to infinity (for a constituent present in the fresh rock and absent in the altered), it is convenient to use a scale in which the divisions become progressively smaller to the right in such a way that the point representing infinity falls at the end of the line. The scale employed in the

text is further modified so that equal percentage losses and gains fall equal distances from the 100 point.

If the number of grams of altered rock actually resulting from the alteration of 100 grams of fresh rock can be indicated by a point on the diagram, it follows that all constituents whose points fall to the right have been lost, and all constituents whose points fall to the left, or inside, have been added during the alteration. Since in general it is not possible to determine how much altered rock results from the alteration of 100 grams of fresh rock, the diagram in general serves to show the *relative* gains and losses, i. e., if any constituent is assumed to have remained constant during the alteration, all constituents to the right of this point represent losses, and all constituents represented by points to the left have been relatively added. If the weight is assumed to remain constant, then all points to the left of "100" line represent gains, all points to the right represent losses. It is important to keep in mind the fact that the diagrams represent changes in terms of the constituents themselves, and not in terms of the total rock mass. Consequently the changes of constituents present in only very small amounts show just as conspicuously as constituents present in considerable amounts, which, obviously, may lead to misinterpretation if this fact is neglected.

Composite "straight-line" diagrams. This type of diagram may be used: (a) in representig the changes shown in a series of analyses of successive stages in a single alteration; (b) in representing simultaneously the changes shown by a number of pairs of analyses of fresh and altered phases.

The manner of representing a series of analyses of successive steps in an alteration is illustrated in Fig. 6, showing the alteration of serpentine rock to iron ore in Cuba. In this type of diagram the analyses (each compared with the original phase) are platted on separate lines, arranged according to depth.

Representation of a number of pairs of analyses of fresh and altered rocks on the same diagram is illustrated by Plate II. In this type of composite diagram the points for each constituent are platted on a separate line and the several points for each pair of analyses connected by lines numbered for purposes of identification. It is convenient for laboratory work to have a printed form which may be used for either of the three types of diagrams described.

Other applications of the "straight-line" diagram. **This method of comparing analyses may be used for comparison of analyses other than those showing metamorphic changes, as, for instance, comparison of igneous rocks; changes in composition of river waters or of mine waters, mineral composition, etc.**

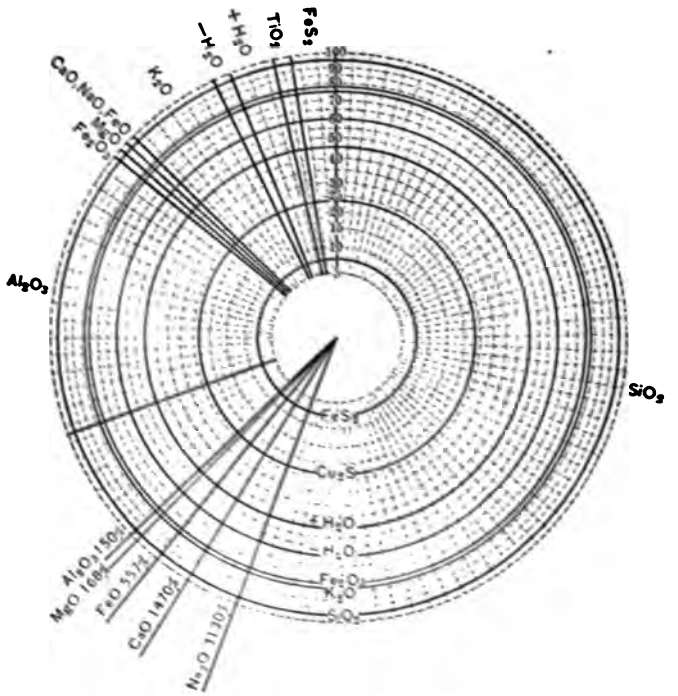

FIG. 26. Circular diagram showing losses and gains of constituents in the alteration of an average acid igneous rock by thermal solutions. After Steidtmann.

Circular Diagrams Representing Gains and Losses of Constituents

A diagram similar in principle to the "straight-line" has been described by Leith and Harder [1] and by Steidtmann.[2] In this

[1] Leith, C. K., and Harder, E. C., The iron ores of the Iron Springs district, southern Utah: Bull. 338, U. S. Geol. Survey, 1908, pp. 28–32.

[2] Steidtmann, Edward, A graphic comparison of the alteration of rocks by weathering with their alteration by hot solutions: Econ. Geol., vol. 3, 1908, pp. 384–385.

diagram, an example of which is shown in Fig. 26, the chemical composition of the altered rock is represented by sectorial divisions of a circle. Values obtained by dividing the percentage of each constituent in the fresh rock by its percentage in the altered rock and multiplying the quotient by 100 are represented to scale by the radii of concentric circles. On assuming a constituent to have remained constant, all constituents whose circles

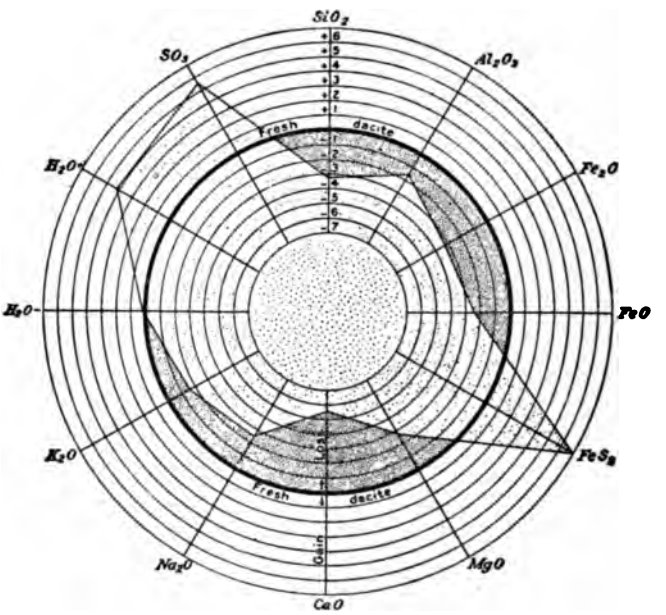

FIG. 27. "Diagram illustrating gains and losses of each chemical constituent in terms of percentage of mass of original fresh dacite. Radial distance between successive circles corresponds to 1 per cent. Gains are outside, and losses inside, of the heavy circle representing fresh dacite. Areas have no significance." After Ransome.

fall outside have been relatively lost and those whose circles fall inside have been gained. The "straight-line" diagram has practically all of the advantages of this type of circular diagram, and being less complicated, has largely superseded it.

Ransome has employed a circular diagram [1] to represent losses

[1] Ransome, F. L., The geology and ore deposits of Goldfield, Nevada: Prof. Paper 66, U. S. Geol. Survey, 1909, p. 181.

and gains of constituents previously calculated from analyses of fresh and altered rocks. This diagram is illustrated in Fig. 27.

The fresh rock is represented by a circle and the losses and gains for each constituent are shown to scale, respectively, inside and outside of this circle, along properly designated radii.

Rectilinear Coördinate Diagram for Comparing Analyses

Rectilinear coördinates have been used in a variety of ways as a means of representing rock analyses and progressive changes in rock alterations. In this type of diagram the constituents may be represented in terms of percentage of total rock mass or in terms of molecular ratios obtained by dividing the percentage for each constituent by its molecular weight.

Fig. 28 illustrates a method employed by Ransome[1] in illustrating progressive alteration of porphyry by vein-forming solutions at Breckenridge, Colorado. The diagram is self-explanatory.

One of the simpliest forms of representing partial analyses of rocks on rectilinear coördinates is the platting of percentages or molecular ratios of two constituents as ordinate and abscissa respectively. This may be varied by platting the percentage or molecular ratio of the single constituent as ordinate and the ratio of percentages or molecular ratios of two other constituents as abscissa.

Iddings[2] has used a form of rectilinear coördinates combined with a modified form of Brögger's diagram (see below) in representing the chemical composition of igneous rocks. In this diagram for each analysis represented, the percentage of silica is platted as the horizontal distance from a zero point and the alkali silica ratio represented by the vertical distance from a zero point. From this point, representing the percentage of silica and the alkali silica ratio, the relative molecular proportions of the six oxides Al_2O_3, K_2O, Na_2O, CaO, MgO, and FeO = (FeO + Fe_2O_3), are

[1] Ransome, F. L., Geology and ore deposits of the Breckenridge district, Colorado: Prof. Paper 75, U. S. Geol. Survey, 1911, p. 97.

[2] Iddings, J. P., Chemical composition of igneous rocks: Prof. Paper 18, U. S. Geol. Survey, 1903, Pl. 1, p. 92.

expressed by radii vector. The ends of these radii vector are connected and the triangular areas thus formed colored, in order to differentiate the several constituents.

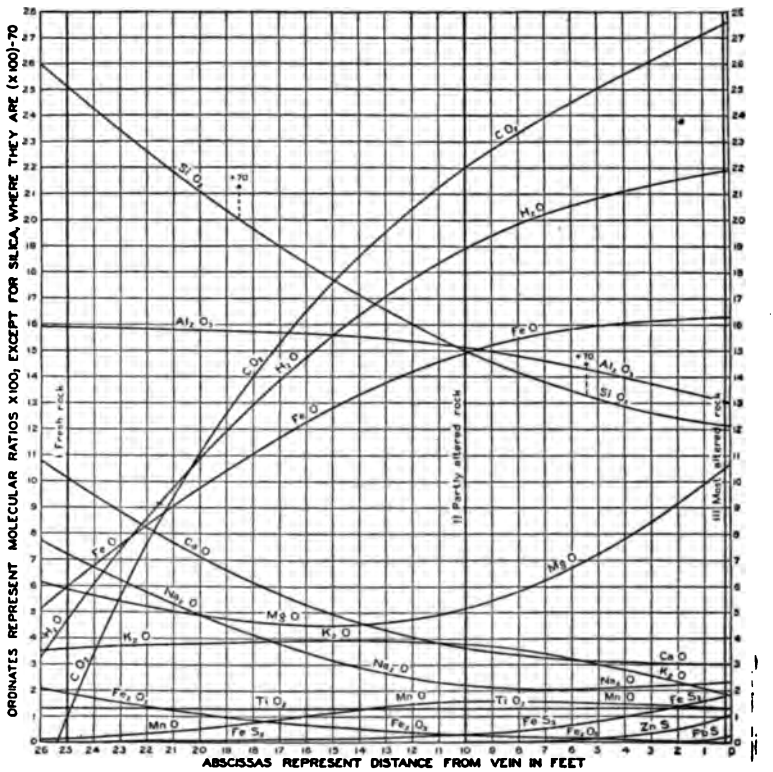

FIG. 28. Diagram showing alteration of diorite prophyry by vein-forming solutions. After Ransome.

Brögger's Diagram for Comparing Analyses

The diagram devised by Brögger,[1] with various modifications, has been widely used. In this diagram the amounts of the eight principal chemical constituents in terms of molecular ratios are

[1] Brögger, W. C., Die Eruptivgesteine des Kristianiagebietes; Das Ganggefolge des Laurdalits, Kristiania, 1898, Pl. I, p. 255.

laid off on radii vector and the intercepts joined by straight lines to form an irregular polygon, the shape of which indicates the general composition of the rock. This is illustrated in Fig. 29 which represents Clarke's average igneous rock.

Columnar Diagrams for Comparing Analyses

The composition of rocks, particularly those representing progressive alterations, may be effectively represented in terms of chemical constituents or in terms of minerals by weight or volume by means of a simple rectangular diagram, in which a rectangle having a long vertical dimension is divided by horizontal lines into areas representing the proportions of constituent, as in Fig. 4.

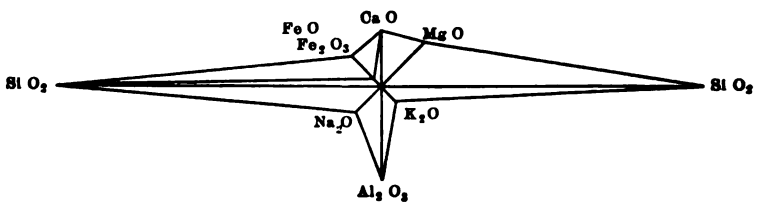

FIG. 29. Clarke's average igneous rock represented by means of Brögger's diagram.

When the volume compositions of rocks are represented, it is effective to represent them by means of an isometric projection of a parallelopiped divided by horizontal planes into blocks proportional to the several constituents. (See Pls. III and IX.)

Representation of Analyses on Triangular Diagram

Trilinear coördinates or "triangular diagrams" may be used to advantage in representing analyses, either chemical or mineralogical, particularly when three of the constituents constitute a large part of the total. In an equilateral triangle the sum of the perpendicular distances from any point in the triangle to the three sides equals the altitude of the triangle. If the altitude of the triangle is divided into 100 units, the sum of the three perpendicular distances from any point to the three sides equals 100, and any ratio of three variables may be expressed in terms of parts per hundred by points in the triangle, each apex being designated as representing respectively 100 parts of each variable.

When the three constituents platted total 100 per cent the analysis is represented by a point, as is illustrated at "A" in Fig. 30. Four constituents are represented at "B," the altitude of the small triangle indicating the percentage of the fourth constituent. Five constituents are represented at "C" and six at "D" by subdividing the small triangles in the same manner as the large primary triangle is divided.

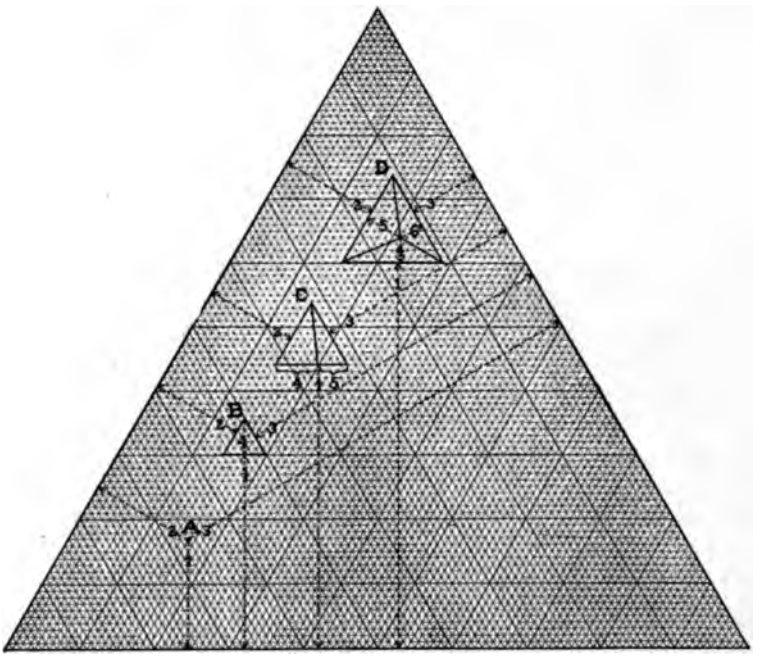

FIG. 30. Triangular diagram illustrating the method of platting three, four, five, and six constituents.

A series of analyses representing a metamorphic change may be represented in this manner, as illustrated by Fig. 5, showing development of bauxite from syenite.

Other uses of triangular diagrams. Trilinear coördinates or triangular diagrams afford a graphical means of solution of certain problems involving three variables.

The problem of combining three analyses to approximate as closely as possible a fourth is described on pages 316–319, in which

the proportions of shales, sandstone, and limestone, which most closely approximate the composition of the average igneous rock, are determined.

A somewhat similar use of the triangular diagram, but one that is rather more complicated, has been made in determining the proportion of average granite and average basalt, which is in best

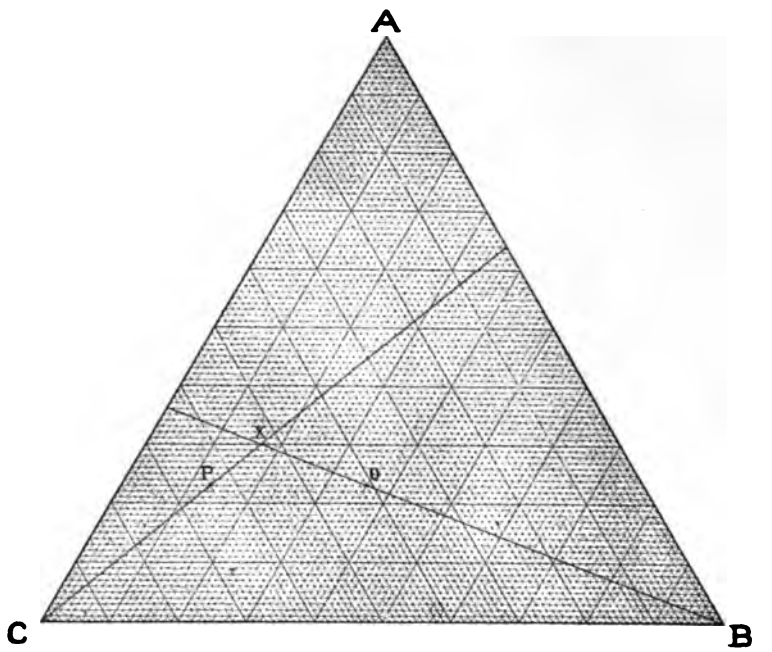

FIG. 31. Illustrating the method of recalculating two, and three quantities to totals of 100 per cent, by means of triangular diagram.

accord with the average composition of shale, sandstone, and limestone. (See pp. 64–67 and reference.)

The triangular diagram serves as a convenient proportional triangle, as a line across it from one corner to any point on the opposite side is a locus of points for which the perpendicular distances to the other two sides are proportional. This serves as a rapid graphical means of calculating two values to a total of 100 per cent. This is illustrated in Fig. 31, in which the point P

COMPARISON OF ANALYSES TO DETERMINE RELATIVE AND ABSOLUTE GAINS AND LOSSES OF CONSTITUENTS

Metamorphic changes in rock in general involve changes in chemical composition due to losses and gains of constituents. These losses and gains can best be studied and determined quantitatively by comparison of analyses representing the rock in different stages of alteration. Reliability of results based on chemical analyses of rocks obviously requires proper field investigation of the rocks to insure that the samples analyzed are really comparable, and requires careful sampling of the rocks to be analyzed.

Chemical analyses represent the relative amounts of the various constituents in terms of weight. Direct comparison of analyses representing different stages in the metamorphism of a rock, can be made only when there has been no change in weight due to the alteration. If 100 grams of fresh rock produce 100 grams of altered rock the losses and gains may be found by direct comparison of the analyses of the fresh and altered rock. When, due to losses or gains of constituents, 100 grams of fresh rock have produced *more* or *less* than 100 grams of altered rock, obviously direct comparison of chemical analyses of the two phases cannot be made. In that case it is necessary to reduce the analyses to a common basis for comparison.

In alterations which have proceeded without change in volume, weight relations between unit volumes of the fresh rock and the altered phases may be calculated, if the mineral specific gravities and the porosities of the two phases are known. Ransome [1] has calculated gains and losses, on the assumption of constant volume, in hydrothermal metamorphism of rocks from Goldfield, Nevada, and from Breckenridge, Colorado.

When weight or volume relations between the two phases are not known, it is possible in some cases to select a constituent which has remained nearly constant in absolute amount during the alteration. With this as a basis the analyses may be recalculated to a comparable basis. (See calculation on basis of constant alumina in weathered Georgia granite, pp. 7–10.)

[1] Ransome, F. L., The geology and ore deposits of Goldfield, Nevada: Prof. Paper 66, U. S. Geol. Survey, 1909, p. 181; Geology and ore deposits of the Breckenridge district, Colorado: Prof. Paper 75 U S Geol. Survey, 1911, p. 96.

jecting the point within the small triangle from the kaolin corner on to the orthoclase-muscovite side. Knowing the ratio of orthoclase to muscovite and the percentage of potash in each, the division of available potash between the two potash minerals may be made and the three minerals calculated.

CALCULATION OF MINERAL COMPOSITION OF A ROCK FROM ITS CHEMICAL ANALYSIS

In metamorphic work it is frequently desirable to translate the chemical analysis of a rock or ore into terms of minerals. The variable and complex composition of some minerals may make an exact translation impossible, but even in these cases it is desirable to express the approximate composition of these minerals in terms of their simpler constituent molecules. The reverse translation, from mineral to chemical composition, while probably less frequently used, is at times of equal importance.

The method most generally employed is to convert the parts by weight into molecular ratios, group these molecular ratios properly to correspond to mineral molecules, and then reconvert the molecular ratios of the minerals into parts by weight. This method is described in detail by Kemp,[1] Iddings and others,[2] and is facilitated by tables of molecular ratios, the most exhaustive being those published by Osann.[3] The computations involved in this method are rather long and laborious even when aided by the tables of molecular ratios.

A more direct method in which the translation is accomplished with less computation by the use of tables of mineral compositions has been published by Harker.[4] Harker's tables, however, do not readily permit of the reverse translation from mineral to chemical composition, and considerable repetition of tables is necessary in order that the percentage of a mineral in a rock may be calculated from any one of the constituent oxides. The principle in-

[1] Kemp, J. F., The recalculation of the chemical analyses of rocks: School of Mines Quarterly, vol. 22, p. 75.

[2] Cross, Whitman, Iddings, J. P., Pirsson L. V., and Washington H. S., Quantitative classification of igneous rocks, 1903.

[3] Osann, A., Beitrage zur Chemischen Petrographie: I. Molekularquotienten zur Berechnung von Gesteinsanalyses, Stuttgart, 1903.

[4] Harker, Alfred, private publication, Cambridge University Press, London, 1910.

volved is illustrated by a calculation from one of Harker's sets of tables.

TABLE XX. SPECIMEN TABLE FOR CALCITE FROM HARKER'S TABLES FOR CALCULATING MINERALS FROM CHEMICAL ANALYSES

Calcite, $CaCO_3$		
CaO	CO_2	Calcite
1	1.274	2.274
2	2.55	4.55
3	3.82	6.82
4	5.10	9.10
5	6.37	11.37
6	7.64	13.64
7	8.92	15.92
8	10.19	18.19
9	11.47	20.47

If a given analysis contains 2.54 per cent lime, in the form of calcite, the per cent of that mineral and of the CO_2 combined in it, in terms of 100 units of rock mass, are obtained from the table as follows:

2.00	CaO	requires	2.55	CO_2	and	makes	4.55	calcite.
.50	"	"	.637	"	"	"	1.137	"
.04	"	"	.051	"	"	"	.091	"
2.54	"	"	3.24	"	"	"	5.78	"

Mineral Slide Rule for Computing Mineral Composition from Chemical Analyses

Another method of calculation, which has the advantage of both speed and accuracy, is by means of a computing diagram which involves the logarithmic principle, illustrated in Fig. 32. It is, in fact, a specialized form of circular slide-rule. As this device has movable parts, it is necessarily represented in a fixed position in the figure and its operation must hence be described rather than illustrated. This device consists of three parts: A heavy white celluloid disc about 8″ in diameter, having a circular logarithmic scale near its margin. Concentric with this logarithmic scale is a smaller movable disc, having a number of equally

spaced concentric circles (one for each mineral). In addition to these two discs there is a radial transparent celluloid arm on the under side of which a fine black line is engraved in line with the center of the circles. This celluloid arm is pivoted to the center of the circles so that the fine black line exactly crosses the center point. The radial arm is free to move about the center and car-

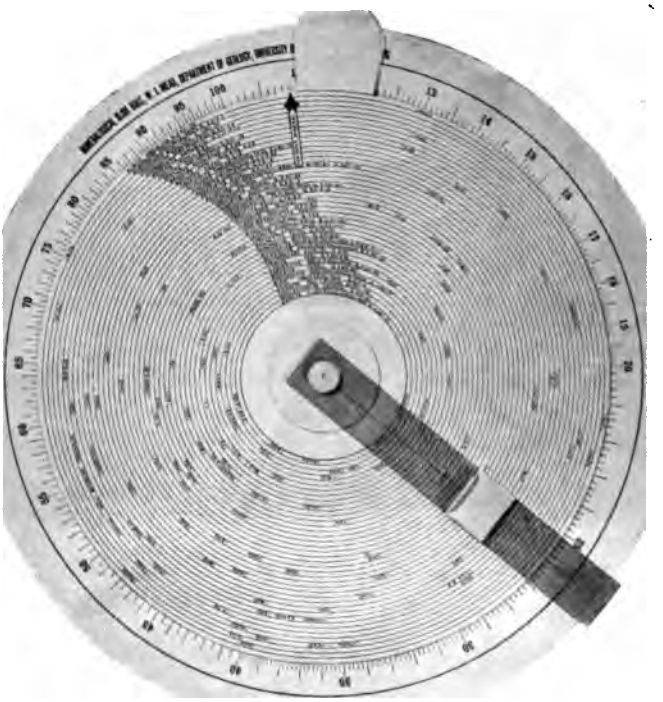

FIG. 32. Mineral slide rule for computing mineral compositions from chemical analyses.

ries a slide which may be set to follow around any one of the concentric circles of the smaller disc. Forty of the common rock-forming minerals, both original and secondary, are included in the diagram and are arranged alphabetically.[1] Following each mineral is its chemical formula and specific gravity.

[1] A similar mineral slide rule for 50 of the important ore minerals has been constructed for laboratory work on the ores.

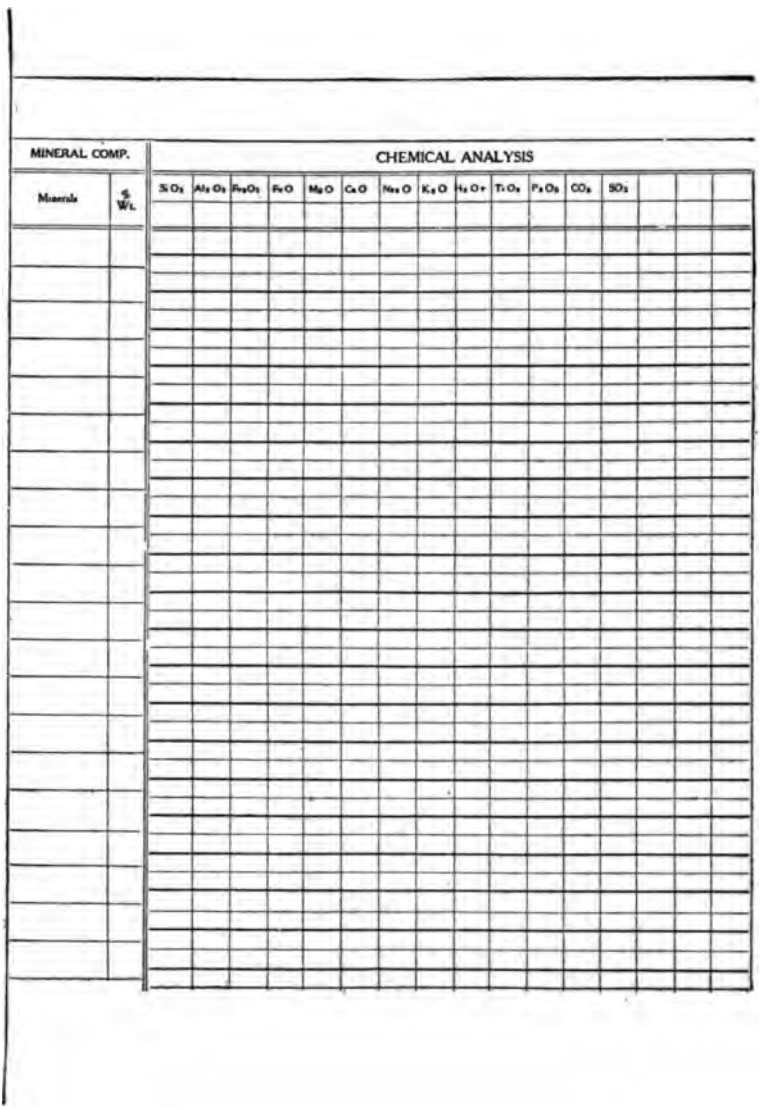

FIG. 33. Form for use with mineral slide rule.

The operation of the mineral slide rule is as follows:

The slide on the radial arm is set to follow the circular scale of the mineral considered. This is accomplished by setting the *outside* edge of the aluminum slide to coincide with the *inner* circle of the scale of the mineral. The celluloid arm is then turned until the fine black line indicates on the logarithmic scale the percentage of one of the constituents of that mineral in the rock. The movable disc carrying the mineral scales is then rotated until the short cross line on the mineral scale, marking the position of the constituent chosen, appears on the black radial line of the celluloid arm. The percentage of the mineral, in terms of total ore or rock, may then be read on the logarithmic scale at the point indicated by the black arrow labeled "per cent mineral." The two celluloid discs are then held immovable and the percentages of each of the constituents of the mineral in terms of total rock are read off the logarithmic scale by moving the celluloid arm until the radial black line coincides successively with each of the constituents marked on the circular mineral scale. The procedure does not differ in principle from that in which the molecular ratios of the constituents are used.

The accuracy of the mineral slide-rule is equivalent to that of a twenty inch slide-rule, or a three to four place logarithm table. It has a higher degree of accuracy than most of the published tables of molecular ratios and does not require interpolation.

In connection with the mineral slide-rule it is convenient to use a printed form, Fig. 33, on which the operations may be tabulated. In this form the chemical analysis is written on a horizontal line at the top of the page, as indicated, and the percentage of oxides used for each of the minerals are in turn subtracted from the total amount in the analysis, so that a balance is each time carried forward. In calculating the chemical analysis from the mineral composition of the rock or ore, the names and percentages of the minerals are written in the proper columns and then the percentage of each oxide (read from the chart) is written in the proper line. On adding the columns for the several chemical constituents, the sums (written at the top) represent the chemical composition of the rock or ore.

CALCULATION OF VOLUME CHANGES

Volume Changes in Alterations of Minerals

The theoretical changes in volume involved in a mineral alteration may be calculated from the chemical equation of the alteration by comparing the molecular volumes of the minerals involved. By way of illustration, consider the alteration of orthoclase to kaolin and quartz, expressed in the following equation:—

	Orthoclase		Kaolin	Quartz
	$K_2O.Al_2O_3.6SiO_2$	$+ 2H_2O + CO_2 =$	$Al_2O_3.2SiO_2.2H_2O$	$+ 4SiO_2 + K_2O.CO_2$
Mol. wt. x No. Mol.	556		258	240
Sp. Gravity	2.57		2.63	2.65
Mol. vol. $= \frac{\text{Mol. wt.}}{\text{Sp. Grav.}}$	216.2		98.1	90.5

98.1 = 54.3 per cent of 216.2
90.5 = 41.8 per cent of 216.2
Volume change considering only kaolin = decrease of 54.7 per cent.
Volume change considering kaolin and quartz = decrease of 12.9 per cent.

Fig. 34 illustrates a printed form useful in the laboratory for tabulating calculations of mineral alterations. The lower portion of the form is used for representing the volume relations graphically.

Volume Changes in Rock Alterations

If the weight relations between an unaltered rock and an altered phase are known, and also the specific gravities and porosities of the two phases, the volume relations may be calculated. The following hypothetical case is given by way of illustration.

One hundred grams of unaltered rock, A, yield 60 grams of altered rock, B. The specific gravity of A is 2.90 and of B 2.60. The porosity of A is 4 per cent and of B 30 per cent.

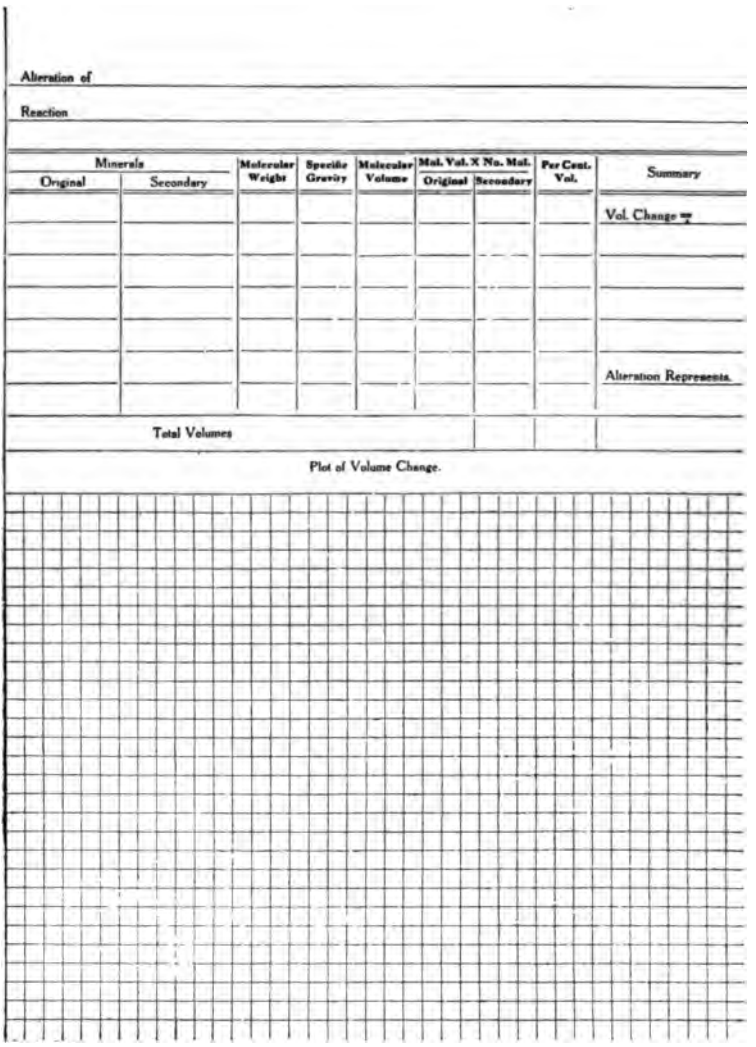

FIG. 34. Form used for tabulating calculations of mineral alterations, and representing volume changes graphically.

Volume of rock in cubic centimeters exclusive of voids =	weight in grams / specific gravity.	
	A.	B.
Volume exclusive of voids in cubic centimeters	34.45	23.8
Volume of voids in cubic centimeters	1.43	10.2
	35.88	34.0

Volume change is decrease of 1.88 cubic centimeters = 5 per cent decrease.

The volume composition of a rock may be expressed in terms of minerals and pore space. The percentages of minerals in a rock may be calculated by dividing the percentage weight of each mineral by its specific gravity and recalculating the quotients to total 100 per cent minus the porosity.

CALCULATION OF ENERGY CHANGES

Quantitative calculation of energy changes involved in mineral alterations is limited by the comparatively few mineral compounds for which the heats of formation have been determined. It is impossible from available data to calculate the net energy change involved in the mineral alterations which take place in the alteration of a rock. It is of interest, however, to perform some of the calculations involving simple chemical changes for which heats of formation are available. For example, we may calculate the energy change in the alteration of ferrous silicate to hematite and quartz, expressed in the following equation:

$$2FeOSiO_2 + O = Fe_2O_3 + 2SiO_2$$

The heats of formation of the compounds involved in gram calories are as follows:

$FeSiO_2 = +254{,}600$. $Fe_2O_3 = +195{,}000$. $SiO_2 = +180{,}000$. $O = 0$

Writing the above equation with the heats of formation below each compound with signs reversed, and balancing the heats of formation, we find that the reaction involves a liberation of 173.4 calories per gram of $FeOSiO_2$.

$$2FeOSiO_2 + O = Fe_2O_3 + 2SiO_2$$
$$-(254{,}600 \times 2) + 0 = -195{,}000 - (180{,}000 \times 2)$$

from which heat liberated = 45,800 calories.

Mol. wt. of $2FeOSiO_2 = 264$.

$$\text{Calories per gram } FeOSiO_2 = \frac{45{,}800}{264} = 173.4$$

In a similar manner the energy change for the alteration of wollastonite to calcite and quartz may be calculated.

$$CaOSiO_2 + CO_2 = CaCO_3 + SiO_2$$
$$-344{,}400 \quad -97{,}650 = -269{,}100 \quad -180{,}000$$

from which heat liberated = 7,050 calories.

$$\text{Calories per gram } CaSiO_3 = \frac{7{,}050}{116} = 60.7$$

For tables of heats of formation see Smithsonian tables and references.[1]

[1] Smithsonian physical tables, 5th revised edition: Smithsonian Institution Misc. collections, vol. 58, No. 1, 1910.

CHAPTER II

SUGGESTIONS FOR LABORATORY STUDY

Laboratory work in metamorphism supplements field observations, which properly include microscopical study of metamorphic rocks, adequate sampling, and may, in many cases, include to advantage porosity and density determinations and concentration and study of heavy residuals. Attention has already been called to the scarcity of quantitative data collected in field work and to the desirability of more emphasis on this phase of metamorphic investigation.

KATAMORPHISM OF IGNEOUS ROCKS BY WEATHERING

Suites of specimens with analyses and thin sections showing gradation from fresh to altered rock are desirable.

Study of Specimens and Thin Sections

Study of specimens for relations of disintegration and decomposition, mechanics of disintegration, and manner in which decomposition works into the rocks and minerals. Note changes in color and hardness of the various minerals during alteration, kaolinization of feldspars, and bleaching and chloritization of ferrous minerals, change in color and appearance of rock.

Where samples of residual clays and soils are available these may be studied microscopically and the unaltered residuals of original minerals separated by washing for examination and study.

Specific gravity and porosity of the several phases should be determined for use in later calculations and for comparison with calculated values. Under the microscope note mineralogical changes and relations of these changes to cleavage and fracture in the minerals. Note relative advancement of alteration in different minerals.

To Compute Changes in Chemical Composition

Write chemical reactions for simple cases of alteration observed microscopically, for instance, orthoclase to clay or sericite.

Develop necessity for reducing analyses to a comparable basis (as, for example, assumption of alumina constant during weathering). Recalculate analyses of altered rock on basis of alumina constant (see pp. 8–9), and determine actual gains and losses of constituents, as shown in Table II, page 8.

Represent losses and gains by means of "straight-line" diagram.

Compare with losses and gains in hydrothermal katamorphism.

To Calculate Mineralogic Changes

From examination of thin sections and specimens, list as completely as possible minerals present in analyzed phases. From chemical analyses calculate mineralogical composition of fresh and altered phases. As a specific illustration, the following Tables XXI and XXII show the successive steps in the calculation of the mineral compositions of the fresh and altered granite, given in Table II, page 8. In these calculations the direct percentage method was used by employing a geological slide-rule. (See descriptions, pp. 299–302.) Similar calculations could of course be made by using the "molecular ratio method."

TABLE XXI. CALCULATION OF MINERALS OF UNALTERED GEORGIA GRANITE FROM ANALYSIS IN TABLE II, PAGE 8

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O+
	68.75	17.59	1.39		.64	3.25	4.54	3.27	.56
Biotite.....5.59	1.95	.85	.54	.76	.64		.18	.48	.19
	66.80	16.74	×	×	×		4.36	2.79	.37
Orthoclase. 16.60	10.75	3.04						2.79	
	56.05	13.70					4.36	×	
Albite 36.90	25.40	7.20				3.25	×		
	30.65	6.50				×			
Anorthite .16.20	7.01	5.95							
	23.64	.55							
Kaolin.....1.39	.65	.55							.19
	22.99	×							.18
Quartz....22.41	22.99								
	×								

TABLE XXII. CALCULATION OF MINERALS OF ALTERED GEORGIA GRANITE, FROM ANALYSIS IN TABLE II, PAGE 8

	SiO_2	Al_2O_3	Fe_2O_3	MgO	CaO	Na_2O	K_2O	H_2O+
	57.04	26.04	1.91	.17	.75	1.91	2.41	9.77
Limonite.....2.23			1.91					.32
			×					9.45
Chlorite.....47	.15	.09		.17				.06
	56.89	25.95		×				9.39
Orthoclase....14.30	9.30	2.62					2.41	
	47.59	23.33					×	
Albite 16.20	11.10	3.14				1.91		
	36.49	20.19				×		
Anorthite 3.73	1.62	1.38			.75			
	34.87	18.81			×			
Kaolin, etc....50.30	22.10	18.81						9.39
	12.77	×						×
Quartz.....12.77	12.77							
	×							

In the calculation of the mineral composition of the unaltered granite, biotite was first calculated (as average biotite given on the slide-rule), using all of the magnesia in this mineral. All iron in the analysis of the fresh granite is given as Fe_2O_3. This is separated into ferrous and ferric iron in accordance with the requirements for biotite. The remaining potash, soda, and lime are calculated respectively into orthoclase, albite, and anorthite molecules. The small remainder of alumina is figured as kaolin and the remaining silica as quartz.

In the calculation of the mineral composition of the altered rock magnesia is figured as chlorite,—soda, lime and potash were thrown into the feldspars, and the remaining alumina, with enough silica to satisfy the alumina-silica ratio of kaolin, together with the combined water, are grouped together under the head of clay minerals, kaolin, etc. In this rock the amount of combined water is in excess of that required for kaolin, and probably indicates the presence of either a more hydrous aluminum silicate, such as halloysite, or, what is equally probable, the presence of hydrous aluminum oxide, such as gibbsite or bauxite. The remaining silica is assumed to be quartz.

To Compute Volume Changes

Translate the calculated mineralogical compositions into terms of volume by dividing the percentage weight of each mineral by its specific gravity. If the percentage weights are assumed to represent weight in grams, dividing by specific gravity gives volume in cubic centimeters. In Table XXIII, page 311, this calculation has been made. By dividing the total weight in grams by the total volume in cubic centimeters, the specific gravity of the rock is obtained.

In the case of the Georgia granite (p. 8) assuming alumina constant, 100 grams of fresh rock yield 67.05 grams of altered rock. This amount of altered rock has a volume of 26.16 cubic centimeters, as compared with a volume of 37.30 cubic centimeters for the minerals of 100 grams of fresh rock, showing a volume difference of 11.14 centimeters or 30 per cent of the original. The difference in volume measures the pore space in the altered rock, caused by leaching of soluble materials. If mechanical slump or settling occurs this porosity is less.

TABLE XXIII. CONVERSION OF PERCENTAGE WEIGHTS OF MINERALS OF FRESH AND ALTERED GEORGIA GRANITE INTO TERMS OF VOLUME

Fresh granite	W Grams per 100	G Specific gravity	Cubic centimeters per 100 grams rock, W/G	
Biotite	5.59	2.9	1.93	
Orthoclase	16.60	2.57	6.46	
Albite	36.90	2.63	14.04	
Anorthite	16.20	2.75	5.89	
Kaolin	1.39	2.63	.53	
Quartz	22.41	2.65	8.45	
Totals	99.09		37.30	
Sp. Gr. of rock = 99.09 ÷ 37.30 = 2.66				
Altered granite				Cubic centimeters in 67.5 grams rock (assuming Al_2O_3 constant)
Limonite	2.23	3.80	.59	.40
Chlorite	.47	2.71	.173	.12
Orthoclase	14.30	2.57	5.56	3.75
Albite	16.20	2.63	6.16	4.15
Anorthite	3.73	2.75	1.36	.92
Kaolin, etc.	50.30	2.50	20.10	13.56
Quartz	12.77	2.65	4.82	3.26
Totals	100.00		38.76	26.16

Specific gravity of altered rock = 100 ÷ 38.76 = 2.58.

To Compute Materials Removed in Solution

It is of interest to consider in terms of minerals the materials lost during alteration of the Georgia granite. In the following table (Table XXIV) these losses have been calculated into terms of minerals, as follows:

TABLE XXIV. CALCULATION OF CONSTITUENTS LOST FROM ALTERATION OF GEORGIA GRANITE INTO TERMS OF MINERALS

	SiO_2	Fe_2O_3	MgO	CaO	Na_2O	K_2O	H_2O+	CO_2	Cl
	30.25	.12	.53	2.74	3.25	1.64	.02	2.74	6.50
Dolomite..2.44			.53	.74				1.17	
			×	2.00					
Calcite....3.57				2.00				1.57	
				×					
Limonite....14		.12					.02		
		×							
Halite 8.27					3.25				5.02
					×				
Sylvite....3.12						1.64			1.48
						×			
Quartz...30.25									
Total..47.79									

Carbon dioxide is added to lime and magnesia to form dolomite and calcite. Water is added to iron oxide to form limonite. Soda and potash are calculated as halite and sylvite, respectively. Silica is figured as quartz. This calculation requires the addition of 2.71 grams of CO_2, 6.50 grams of chlorine, making a total of 47.79 grams of soluble products from 100 grams of fresh granite. Considering both the residual material and the soluble material, 100 grams of fresh granite have increased to a total of 113.29 grams of alteration product. These minerals are converted into terms of volume in Table XXV, p. 313.

TABLE XXV. CALCULATION OF MINERALS IN TABLE XXIV INTO TERMS OF VOLUME

Materials removed in solution	W Grams per 100 grams fresh granite	G Specific gravity	W/G Cubic centimeters per 100 grams fresh granite
Dolomite	2.44	2.85	.857
Calcite	3.57	2.72	1.31
Limonite	.14	3.8	.04
Halite	8.27	2.13	3.88
Sylvite	3.12	1.95	1.60
Quartz	30.25	2.65	11.40
Total	47.79		19.087

$$\text{Av. Sp. Gr.} = \frac{47.79}{19.087} = 2.50$$

Represent graphically (see Plate III) the volume compositions of the fresh granite and of the alteration product on a comparable basis, assuming alumina to have remained constant.

HYDROTHERMAL KATAMORPHISM OF IGNEOUS ROCKS.

(See Chapters IV, Part I, IV, Part II)

Study specimens macroscopically for physical and mineralogical changes. Note changes in color, hardness, texture, etc. For use in future calculations determine specific gravity and porosity of specimens. Distinguish, if possible, hydrothermal alteration from later katamorphic alteration by cold solutions, which is commonly found superimposed upon earlier hydrothermal alterations.

In thin sections note the manner of alteration of the minerals. So far as possible, write chemical reactions for the observed mineral alterations and make at least tentative inferences as to the nature of the solutions accomplishing the alteration. Calculate volume changes for the observed mineral changes.

Represent losses and gains of constituents by means of "straight-line" diagram. A circular diagram used by Ransome [1] in representing graphically the changes in composition in hydrothermal

[1] Ransome, F. L., The geology and ore deposits of Goldfield, Nevada: Prof. Paper 66, U. S. Geol. Survey, 1909, p. 181.

alteration is also effective. Compare with losses and gains in weathering of igneous rocks. In the case of hydrothermal alterations (see Pls. V, VI, XIV, XV) it has been seen that alumina behaves in a rather erratic manner, that is, it may be either gained or lost and probably does not remain constant. It cannot generally be used as a constant against which to measure the losses and gains of other constituents, as in weathering. It is then necessary, before conclusions can be made as to absolute gains and losses, to determine in some way the weight relations between the fresh and altered rock. If the alteration has proceeded with constant volume, the weight relations are in direct proportion to the specific gravity of the fresh and altered rock, after correcting actual rock volume for differences in porosity of the fresh and altered phases. When the number of grams of altered rock produced by 100 grams of fresh rock have been determined, this point may be indicated on the "straight-line" diagram and the gains and losses noted with reference to it. When the weight relationship between the fresh and altered phases cannot be determined, either on the basis of inferred volume relation or by calculation from a constituent which has remained nearly constant during the alteration, the *relative* gains and losses only are indicated on the "straight-line" diagram.

Calculation of Mineralogical Changes

Calculate the mineral compositions of the phases of the alteration for which analyses are available in terms of the minerals observed in thin section. It is desirable to represent graphically by means of columnar diagrams (see Fig. 4) or similar device the mineralogical compositions of the analyzed phases in terms of weight and in terms of volume.

Comparison of Hydrothermal Katamorphism and Weathering of Igneous Rocks

Compare the physical, chemical, and mineralogical changes in hydrothermal metamorphism of igneous rocks with those of weathering of similar rocks. (See pp. 46–47.) This subject has been discussed in some detail by Steidtmann.[1]

[1] Steidtmann, Edward, A graphic comparison of the alteration of rocks by weathering with their alteration by hot solutions: Econ. Geol., vol. 3, 1908, pp. 381–409.

KATAMORPHISM OF SEDIMENTARY ROCKS (SEE CHAPTER V, PART I)

The suggestions for laboratory study of katamorphism by weathering of igneous rocks may be applied with slight modification to the study of the katamorphism of the sediments. The products of katamorphism of sediments are generally incoherent soils or residual clays, and the determination of porosity and specific gravity entails certain difficulties not met with in solid rock specimens. (See p. 286.)

KATAMORPHISM OF ORES

This is a phase of metamorphism which offers practically unlimited field for quantitative work in the laboratory. Methods used in the study of the katamorphism of igneous and sedimentary rocks may in general be applied.

Ores Developed as Residual Products of Katamorphism

The Lake Superior iron ores, lateritic iron ores of eastern Cuba, and the Arkansas bauxites are typical of this type of alteration. The Lake Superior iron ores are concentrated from a rock consisting essentially of hematite and quartz by the removal of quartz in solution. For a quantitative discussion of the origin of these ores, with analyses and other data, see Monograph 52, U. S. Geological Survey, Chapter XVII. The Cuban iron ores have been described in detail (see pp. 38–42, and references). This development is simply a case of katamorphism by weathering of a basic igneous rock. The development of the bauxite deposits of Arkansas by weathering of syenite is also illustrative of the katamorphism of ores. (See pp. 25–37, and references.)

Ores Concentrated Largely by Solution and Redeposition

The changes involved in the katamorphism of sulphide deposits afford an interesting field for quantitative study. When chemical analyses and other data for the several zones are available, the changes in density and porosity, in chemical composition, and in mineralogical composition may be studied quantitatively. The disseminated porphyry deposits of the Southwest offer attractive

fields for this type of investigation. On pages 56–58 certain calculations have been made from limited data from a drill hole in one of these districts. Complete chemical analyses, showing gradation from the leached upper zone through the oxide zone, secondary sulphide zone and into the primary sulphide zone in one of these porphyry copper deposits, would afford opportunity for working out all of the changes involved in the alteration of the ore and rock in a manner similar to that employed for the Cuban ores and for the Arkansas bauxite.

Study of the katamorphism of ores should, where possible, be supplemented and correlated with the composition of the underground waters associated with the ores. A large amount of information on secondary sulphide enrichment and the composition of mine waters has been brought together by Emmons.[1]

TO CALCULATE PROPORTIONS OF SEDIMENTS DERIVED BY KATAMORPHIC REDISTRIBUTION OF AVERAGE IGNEOUS ROCKS

In Chapter VII, Part I, the redistribution of the constituents of igneous rocks by katamorphism has been discussed in some detail. Reference in this chapter is made to calculations of the relative amounts of shale, sandstone, and limestone, produced by katamorphic redistribution of the average igneous rock. The general problem involved is the combination of three analyses in such a manner as to equal as closely as possible a fourth. The method of solution described in the following pages is applicable to any problem of this type.[2]

Briefly stated, the problem is to determine in what proportions the average shale, sandstone, and limestone analyses should be combined to approximate most closely the average composition of the igneous rocks. Most of the mass of the sedimentary rocks has been ultimately derived from the igneous rocks, but in the process of redistribution water and gases from the hydrosphere

[1] Emmons, W. H., The enrichment of sulphide ores: Bull. 529, U. S. Geol. Survey, 1913.

[2] This general problem is met with in a number of special problems, as for instance, the proportion of shale, sandstone, and limestone produced by the katamorphic redistribution of a specific rock area, or to cite a very special problem, the proportions of ore, flux, and fuel required in a furnace mixture to yield a slag of a certain desired composition.

and atmosphere combine with constituents derived from the igneous rocks, and enter into the constitution of the sediments. These elements, which are not derived from the igneous rocks, must be omitted from the analyses of the sediments before any quantitative estimates may be made. In Table XXVI the average analyses of the igneous rocks and sediments have been recalculated, omitting the constituents not derived from the igneous rocks.

TABLE XXVI. AVERAGE IGNEOUS AND SEDIMENTARY ROCKS RECALCULATED, H_2O, CO_2, SO_4 FREE, FROM ANALYSES IN TABLE VI, PAGE 63

	Average igneous rock	Average shale	Average sandstone	Average limestone
SiO_2	61.82	63.96	84.05	9.02
Al_2O_3	15.47	16.95	5.10	1.40
Fe_2O_3	2.66	4.41	1.15	.94
FeO	3.53	2.69	.32	
MgO	3.98	2.68	1.25	13.72
CaO	4.93	3.42	5.89	74.09
Na_2O	3.51	1.43	.48	.09
K_2O	3.08	3.56	1.41	.57
TiO_2	.76	.72	.27	.10
P_2O_5	.27	.18	.08	.07
Totals ...	100.00	100.00	100.00	100.00
Totals before recalculation		92.13	93.57	57.69

The problem may be solved by graphic methods as follows:

In Fig. 35 any point in the triangle represents, in parts per hundred, some ratio of shale, sandstone, and limestone. These percentages may be read directly from the trilinear coördinates which parallel the sides of the triangle. In Table XXVI the average igneous rock contains 61.82 per cent silica, the average shale 63.96 per cent, the average sandstone 84.05 per cent, and the average limestone 9.02 per cent. It is evident, since the average igneous rock contains more silica than does the limestone and less than either of the two other sediments, that some combination of limestone and shale could be made which would contain the same per-

centage of silica as the average igneous rock, and, likewise, that some combination of limestone and sandstone would yield the same result. The ratio in which limestone and shale combine to yield a combination containing 61.82 per cent silica may be found

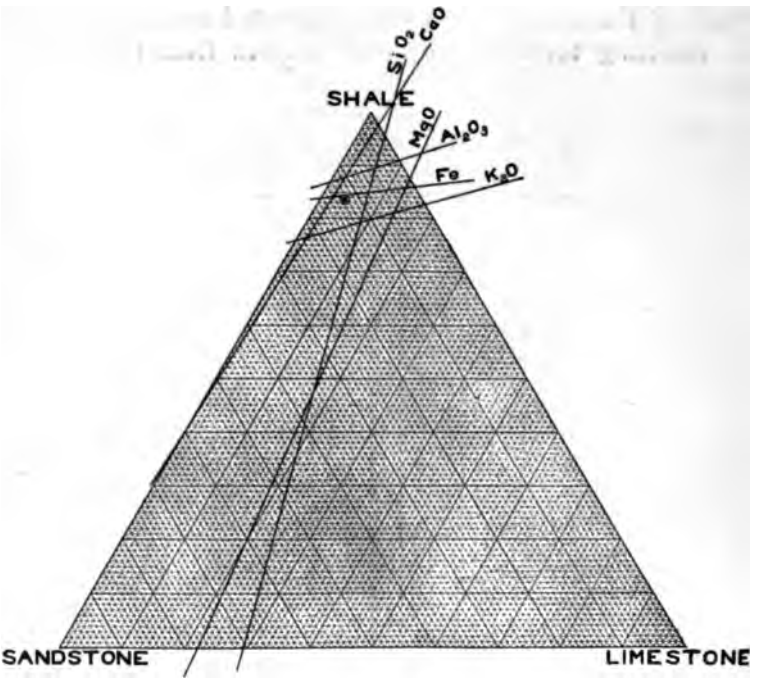

FIG. 35. Graphic determination of proportions of shale, sandstone, and limestone produced by katamorphic redistribution of average igneous rock.

as follows: Let X equal the percentage of shale entering into the combination, and Y the percentage of limestone.

$$X + Y = 1$$
$$63.96\,X + 9.02\,Y = 61.82.$$

Solving those equations for X and Y, Y equals .96, and X equals .04. This ratio may be expressed on the side of the triangle (Fig. 35), connecting the shale and limestone corners by a point four divisions from the shale corner. In a similar manner it is found that the combination of 29.5 parts of limestone and 70.5 parts of sandstone yields an average silica content equal to that

of the average igneous rock. This ratio may be expressed by a point on the sandstone-limestone side of the triangle. A line across the triangle connecting these two points includes all points representing possible combinations of shale, sandstone, and limestone equal in silica content to the average igneous rock. In a similar manner lines are drawn on the triangle for each of the constituent oxides, excepting the oxides of iron, this element being considered without oxygen. (See Fig. 35.) If the problem permitted an exact solution, the lines for the several oxides would intersect at a common point, i. e., some definite ratio of shale, sandstone, and limestone would yield an average analysis exactly like that of the average igneous rock. Since this is not the case, the lines on the triangle do not intersect at a common point. The comparatively close grouping of intersections, however, makes it possible to select some point which will represent approximately a composite of the sediments closest to the average igneous rock in composition.

The point which best satisfies the requirements expresses a ratio of 83.5 parts shale, 12.5 parts sandstone and 4 parts limestone.

From Table XXVI it is seen that the "derived" elements of the shales make up only 92.13 per cent of the mass, of the sandstone only 93.57 per cent, and of the limestone only 57.69 per cent. It is evident, then, that in the above ratio, 83.5 represents only 92.17 per cent of the total mass of the shales, 12.5 represents only 93.56 per cent of the total mass of the sandstone, and 4 represents only 57.86 per cent of the total mass of the limestone. If these figures are modified to take into account the entire mass of sediments, there results a ratio of shales, sandstones, and limestones of 81.7:12:05:6.25, or in even numbers 82:12:6.

Regardless of the exact accuracy of this proportion, it is obvious from the position of the lines on the triangle that shale greatly predominates over the other sediments.

ANAMORPHISM OF CLAYS

Cementation or Induration of Clays and Muds into Shales

In laboratory study of clays it is desirable to compare the physical properties, such as color, texture, plasticity, shrinkage on air

drying, etc., of clays of various origin; for example, residual clays from basic igneous rocks, from acid igneous rocks, from limestone, lacustrine clays, fluviatile clays, glacial clays, and marine clays. Under the microscope their mineralogical compositions and microtextures should be studied.

The subject of the plasticity of clay and its causes offers an interesting field for experimental work.

The textures and mineralogical compositions of shales may be studied in thin sections and also in powdered samples under the microscope and compared with the composition and texture of unconsolidated clays.

When it is difficult to secure actual suites of specimens showing gradation from unconsolidated clay to shale it is of interest to compare chemical analyses of clays in general with those of shales. If a large number of comparisons are made, or better, if averages of large numbers of clay analyses are compared with averages of similar numbers of shale analyses, the differences should be of some significance. In general, it has been found that the unconsolidated clays contain more combined water than the shales.

The subject of the mineralogical composition of clays and of shales and of the mineralogical changes which accompany the consolidation of clays has received a comparatively small amount of attention by investigators. The fine-grained nature of these materials makes their mineralogical study rather difficult, but not impossible if sufficiently high magnifications are used. Reference has been made elsewhere (p. 105) to certain work by Hutchings, and the reader is referred to this paper for discussion of his rather elaborate investigation of this subject, in which certain laboratory methods for the study of these fine-grained clastics are described. It is desirable in the laboratory to attempt to calculate the mineral composition of shales and clays from chemical analyses. This obviously involves many difficulties, as the mineralogy of the rocks is only imperfectly understood. The adsorption of alkalies and of other materials, such as phosphoric acid and titanic acid, by clays, is a subject of importance which merits attention in the laboratory.

On page 76 the average mineral composition of a shale, calculated from the average analysis in the table on page 63, is given. This average mineral composition was calculated from the analysis

as follows: P_2O_5 and SO_3 were combined with sufficient lime to form apatite and gypsum respectively. The remaining lime, carbon dioxide, and part of the magnesia form calcium and magnesium carbonates. Ferrous iron and magnesia, with part of the alumina, silica and water, were combined to form chlorite. The formula for chlorite used was $4H_2O.5(MgO,FeO).Al_2O_3.3SiO_2$. The ferric iron, with sufficient water, was calculated as limonite. It is possible that at least part of the ferric iron is present as a hydrous ferric silicate, nontronite. The remaining alumina, soda, potash, and water were distributed into orthoclase, albite, paragonite, muscovite and kaolin molecules, solving for the relative amounts of feldspars, mica, and kaolin, by means of simultaneous equations, and adjusting emperically the relative amounts of orthoclase and albite. The remaining silica represents free quartz.

By what processes is the cementation of clay into shale accomplished? What constituents, if any, have been introduced during cementation? What constituents have been lost during the process?

Dynamic Metamorphism of Shales

Specimens of shale, slate, and pelite schist, together with thin sections and analyses, are desirable for study of dynamic metamorphism of pelites. If careful search were made in the proper places it should be possible to find actual suites of specimens showing gradation from shale to slate or from slate to schist in the same formation. To the writers' knowledge no actual series of this kind have been described or analyzed. It is indeed strange that this phase of metamorphism of the most abundant class of sediments should not be represented by a single reliable gradational series, or a single comparable pair of analyses. However, much can be obtained from a comparative study of slates, shales, and pelite schists in general. In thin sections under the microscope the mineralogical compositions of these types should be studied in detail and compared.

Write chemical equations and calculate volume changes for observed mineral alterations. Note relation of rock cleavage to mineral cleavage and orientation. (See pp. 171–172.) Note the evidence for the recrystallization of original minerals, development of new minerals from primary materials, behavior of harder mineral

individuals such as quartz and feldspar. What evidence is there in the rocks and thin sections to indicate that the dynamic metamorphism was developed by pressure and rock movement?

Compare average analyses of the various phases of the series by means of "straight-line" diagram. Calculate mineralogical composition from these analyses and make comparison. Which minerals develop most abundantly in the schists? What rock constituents are required for these minerals? What rock constituents are not required?

Contact Metamorphism of Shales

Geological literature contains a considerable number of descriptions of alteration of shale at igneous contacts. Numerous pairs or series of chemical analyses showing gradation or alteration at the contact are available. An excellent series is one published by Hawes,[1] describing the alteration of shale at the granite contact at Mount Willard, New Hampshire. Suites of specimens, thin sections, and analyses should be available for laboratory work. Suggestions for study in previous types of metamorphism apply equally well. The laboratory study should include investigation of physical, chemical, and mineralogical changes involved in the alteration. Mineralogical compositions may be determined in thin sections and calculated from a chemical analysis. The losses and gains of constituents involved in alteration may be represented by means of the "straight-line" diagram or some other graphical device. Chemical reactions should be written for the observed mineral alteration and volume changes calculated. Volume changes for the rock, as a whole, should be calculated from the physical determinations of the specimens and the calculated mineral compositions.

What elements and minerals have been gained by contact metamorphism? What is evidence for introduction of materials from magmatic sources?

ANAMORPHISM OF SANDS

Laboratory work on the anamorphism of sand may well be preceded by preliminary examination of sands in general and study of their probable mode of origin.

[1] Hawes, G. W., Albany granite and its contact phenomena: Am. Jour. Sci., vol. 21, 1881, p. 28.

Cementation of Sands

Study in thin sections and hand specimens the various kinds of cementing materials in sandstone. The principal cements found will be quartz, carbonate cements, and iron oxides. Which two are most often associated? In what way do calcareous cements bind the particles together. In what manner does iron oxide act as a cement? Note that quartz cement is present in three different forms, as finely-divided individuals or grains filling interstices, as crystalline quartz deposited between the grains but bearing no relation to the crystallographic directions of the original grains, and as crystalline quartz in the form of enlargements and in crystallographic continuity to the original grain. (See p. 123 and Fig. 12, p. 123.) Note enlargements evidenced by glistening faces in sandstone specimens. In many sandstones the quartz enlargements are very apparent under the hand lens, appearing as the faces of small quartz crystals which have been built on to the original rounded grains.

Study and compare thin sections of quartzite and sandstone. What distinctions can you make between cementing materials of quartzite and of sandstone?

Note evidence of recrystallization of quartz as shown by loss of clastic texture and increase in size of grains. (See pp. 123–124.)

How much silica is required per unit volume of average sandstone to accomplish complete cementation? What is the source of this silica?

Dynamic Metamorphism of Sands.
(See Plate XI, Chapter II, Part II)

Gradations from quartzite to sericite-quartz-schist have been observed in the field and analyzed and described. Study thin sections and hand specimens of gradation series if available, noting change in texture and relative abundance of minerals. Note manner of development of sericite and its relation to the cleavage of the schist.

Compare chemical analyses of the quartzite and schist, representing losses and gains of constituents graphically by "straight-line" diagram, or other means. Calculate mineral compositions of the phases from analyses and represent graphically losses and

gains of minerals in terms of weight and volume. Represent volume change involved in the development of the schist from the quartzite. This will necessitate establishing the weight relations between the quartzite and schist, that is, the number of grams of schist developed from 100 grams of quartzite. This involves the necessity of assuming that some constituent remained constant during the alteration or determining weight relations on some other basis. See Plate XI and list of analyses. What substances have been introduced? What lost?

Contact Metamorphism of Sands

(See pp. 127, 131)

See description of contact altered sandstone by W. S. Bayley [1] on Pigeon Point, Minnesota; also description of alteration of quartzite and sandstone by Moyie Sill, by R. A. Daly.[2] Laboratory work should include examination of specimens and thin sections, study of the mineralogical changes shown in thin sections, and examination and study of the changes involved in the development of the peculiar spotted structure typical of contact altered quartzite. Calculate losses and gains of chemical constituents from analyses, calculate mineral compositions of various phases from analyses, determine volume changes involved in individual mineral alterations and in the alteration of the rock as a whole.

ANAMORPHISM OF CARBONATES

Cementation or Induration of Carbonates

Compare macroscopically and microscopically marls and calcareous oozes with limestones. What has been the principal method of solidification? Study thin sections of limestones in general, noting different degrees of crystallization. Study relations of calcite and dolomite on smooth surfaces of limestone by staining.[3]

[1] Bayley, W. S., The eruptive and sedimentary rocks on Pigeon Point, Minnesota, and their contact phenomena: Bull. 109, U. S. Geol. Survey, 1893.

[2] Daly, R. A., Geology of the North American Cordillera at the Forty-ninth parallel: Mem. 38, Geol. Survey Can., pt. 1, 1912, pp. 226–252.

[3] Johannsen, A., Manual of petrographic methods: New York, 1914, p. 565.

Dynamic Metamorphism of Carbonates

Pure limestones rarely develop schistose structure. Why? Study in hand specimens and thin sections coarsely crystalline marble from regions which have undergone intense dynamic metamorphism.

Impure limestones, or dolomitic limestones develop schists by dynamic metamorphism. (See pp. 136–137.) In gradational series of samples and analyses, if these are available, chemical and mineralogical changes should be studied as described in the foregoing suggestions for laboratory study of metamorphic changes. Dynamic metamorphism of dolomites develops talc schists. Write the chemical reaction for alteration of dolomite to talc. Calculate volume changes involved.

Contact Metamorphism of Carbonates

(See pp. 137–138)

The frequent association of metalliferous deposits with contact zones of carbonates has called attention to this type of alteration and many instances have been described in terms of chemical and mineralogical changes. (See Plate XII, and references.)

One of the principal problems involved in study of the contact metamorphism of limestones is the question of the source of the constituents of the silicates in the altered phases. (See pp. 140–145.) The matter of volume change during contact metamorphism is one of the essential considerations in connection with this problem. If the limestones have remained constant in volume during alteration, it follows that the silicates of the contact phase have metasomatically replaced original limestone constituents, being introduced into the limestone from outside sources. On the other hand, if the alteration has been accompanied by marked decrease in volume, the silicates may be the original impurities of the limestone concentrated by removal of the carbonates. Field and laboratory study of the question of volume change is essential to the solution of this problem. Laboratory work with specimens, thin sections, and analyses should be carried on with this problem in mind.

IDENTIFICATION OF SCHISTS AND GNEISSES

The problems involved in the determination of schists and gneisses have been discussed on pages 222–242. This subject presents a large and inviting field to the student in the laboratory. Mineralogical and textural features of schists of known sedimentary origin and of known igneous origin should be tabulated and compared. Chemical analyses of schists of known sedimentary and igneous origin should likewise be studied and compared.

The heavy accessory minerals, such as zircon, apatite, etc., may be concentrated by crushing and panning, as previously described, from specimens of schists of known origin, both igneous and sedimentary. See discussion by Trueman [1] on the use of zircon as a criterion for the identification of the origin of foliated rocks. Work on schists and gneisses of unknown origin would naturally consist in comparing criteria or differences obtained from schists of known origin.

[1] Trueman, J. D., The value of certain criteria for the determination of the origin of foliated crystalline rocks: Jour. Geol., vol. 20, 1912, p. 244 et seq.

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